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RELATIVE DURATIONS OF THE TERTIARY SERIES AND STAGES IN NEW ZEALAND

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Summary

A stratigraphic column showing relative lengths of the Tertiary series and stages of New Zealand is presented. Estimates of the relative durations were derived through comparisons of estimates made by five leading New Zealand paleontologist-stratigraphers based upon their concepts of evolutionary changes, thickness of sedimentary sections, and distribution of rock units.

INTRODUCTION

Recently, in an attempt to gain some appreciation of the relative proportions of the series and stage units of the Tertiary as utilised in New Zealand, five leading paleontologist-stratigraphers were questioned. Dr C. A. Fleming and Mr N. de B. Hornibrook, New Zealand Geological Survey, Dr J. M. Marwick, sometime Senior Paleontologist, New Zealand Geological Survey, Mr P. Vella and Dr H. W. Wellman, Department of Geology, Victoria University of Wellington, participated and provided the data here presented. The results of the poll are of such general interest, and of such uniformity, that it would seem useful to present the data for more widespread review. It must be emphasised that the results given here are expressed in relative terms only and no attempt has been made to equate them to an absolute time scale.

In the absence for the foreseeable future of an absolute time scale based upon radioactivity methods, the method discussed below was attempted. The preliminary results obtained through potassium-argon ratios in glauconitic sands presented by Lipson (1958) are somewhat contradictory. From the

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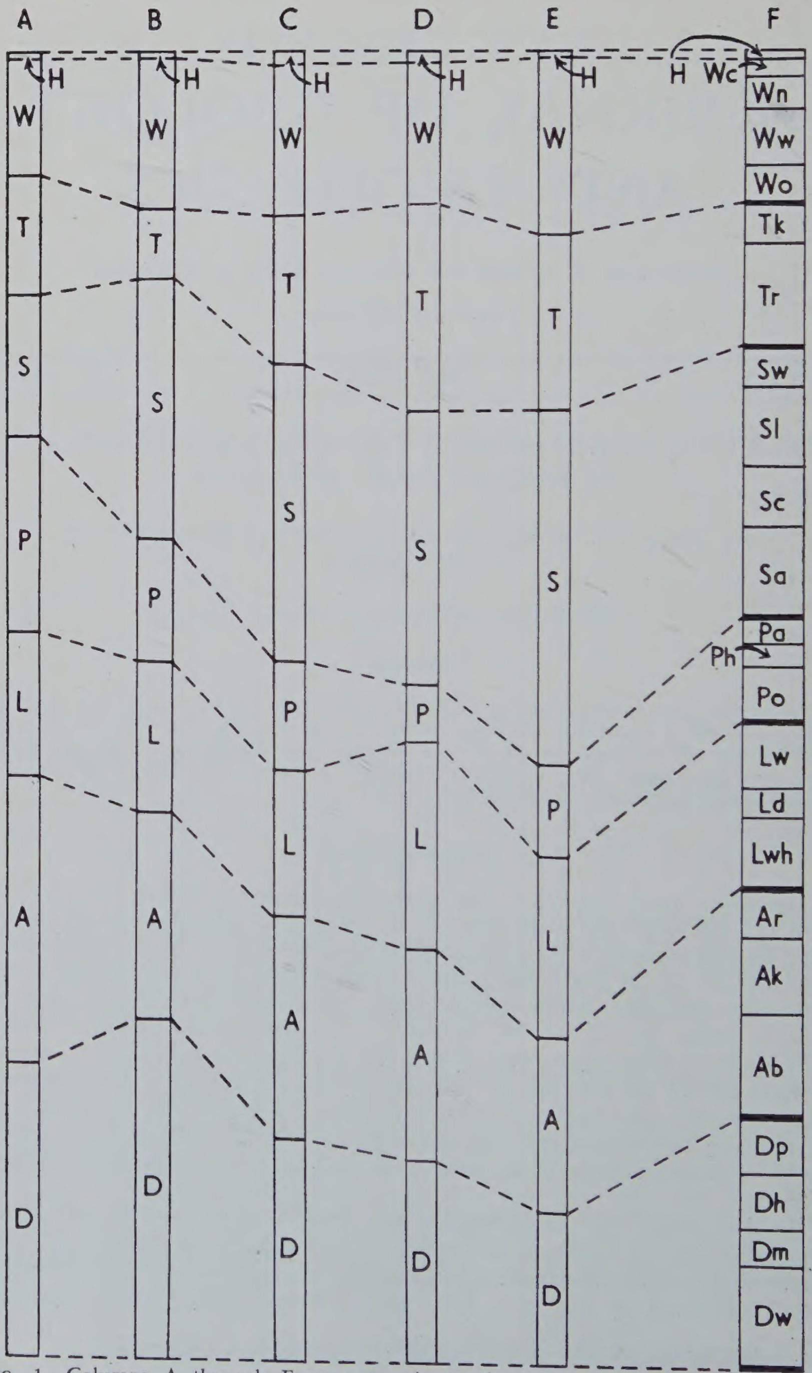

FIG. 1—Columns A through E present estimates by members of the panel of the proportional lengths of series. Column F is a summary of opinions, with series and stage durations represented by mean values obtained from panel estimates.

earliest stages of geological thought man has been concerned with the length of earth history or of the proportional lengths of episodes thereof. In attempting to gain this information, rates of various geological processes, rates of evolution, thicknesses of sedimentary sections have been utilised as the basis for estimates. Although most of these estimates have been inadequate in delimiting the expanse of geological time, the proportions suggested have been strikingly accurate in the light of more modern techniques. A detailed review of these attempts is given by Zeuner (1950). Two of the most utilised proportional scales of geological time are based upon data obtained in this fashion inserted into the framework provided by radioactivity methods. The well known "A" and "B" scales of Holmes (1947) are largely drawn from estimates of thicknesses of sedimentary sections; while Simpson's (1944) proportions are fundamentally those of Matthew (1914) derived from analysis of rates of horse evolution.

New Zealand's Tertiary section, remarkably complete and finely subdivided (Finlay & Marwick, 1947), should provide an adequate record upon which to base such estimates. Accordingly a panel consisting of the above mentioned geologists was asked to prepare estimates based upon their knowledge of the Tertiary record:

Dr C. A. Fleming – Macropaleontology and Stratigraphy.
Mr N. de Hornibrook – Micropaleontology and Stratigraphy.
Mr P. Vella – Micropaleontology and Stratigraphy.
Dr J. M. Marwick – Macropaleontology (as recorded in magnitude of faunal changes).
Dr H. W. Wellman – Stratigraphy (as recorded in thicknesses of sedimentary sections).

After careful analysis of the opinions given by the members of the panel, it is felt that the estimates are mutually complementary in the spheres of individual interest and activity and present an adequate basis for a relative stratigraphic scale.

METHOD AND RESULTS

The panel was asked to independently prepare estimates of: (1) stage lengths; and (2) of the series lengths. The former were made proportional within each of the series by assigning to each of the series a duration of 12 units, these to be apportioned to the component stages. The results of this endeavour were remarkably uniform indicating a high degree of concurrence of thought. To avoid utilisation of an absolute time scale in preparing the estimates series lengths were necessarily more deviously obtained. The assumption was made that the Hawera Series is the shortest, and accordingly it was assigned a unit length. The panel then estimated the duration of the other series relative to the Hawera. As would be expected, the variation obtained was of a somewhat greater magnitude than that for the stage duration estimates. These data were then recalculated as a percentage of total Tertiary time (excluding the bulk of the Paleocene). This is the basis for the columns presented in fig. 1. Table 1 gives the extremes of the estimates as recalculated in percentages. It should be noted that the apparent discrepancies are to a large extent caused by variation in series duration estimates.

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Table I—Range and Mean of Estimates of Duration of Series and Stages Expressed as Percentages of Total Tertiary Time

	Minimum	Maximum	Weighted Mean
Hawera	0·2	1·0	1·0
Wanganui	8·9	13·5	11·0
Castlecliffian	0·9	2·7	2·0
Nukumaruan	1·8	2·7	2·0
Waitotaran	3·6	4·6	4·0
Opoitian	1·8	3·6	3·0
Taranakian	5·8	15·8	11·0
Kapitean	1·8	3·6	3·0
Tongaporutuan	7·3	9·1	8·0
Southland	11·1	27·0	20·5
Waiauan	1·7	3·4	3·0
Lillburnian	3·4	10·3	6·0
Clifdenian	1·7	6·8	4·5
Altonian	6·8	6·8	7·0
Pareora	4·1	17·8	8·0
Awamoan	1·3	4·0	2·0
Hutchinsonian	0·6	2·0	2·0
Otaian	2·7	6·0	4·0
Landon	11·1	15·8	12·5
Waitakian	4·2	6·2	5·0
Duntroonian	1·0	4·2	2·5
Whaingaroan	3·1	6·2	5·0
Arnold	13·5	22·2	17·0
Runangan	2·8	4·2	3·5
Kaiatan	4·2	7·0	5·0
Bortonian	7·0	8·4	8·5
Dannevirke	13·5	25·9	19·0
Porangan	2·2*	6·3*	5·0
Heretaungan	2·2*	6·3*	5·0
Mangaorapan	2·2*	2·2*	2·5
Waipawan	6·3*	6·3*	6·5

*Estimates submitted only by Hornibrook and Vella.

A MINERALOGICAL STUDY OF SOIL FORMATION IN FOUR RHYOLITE-DERIVED SOILS FROM NEW ZEALAND*

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Summary

In four yellow-brown earth and podzolised yellow-brown earth soils derived from rhyolite, much evidence concerning mineralogical changes attendant upon weathering and podzolisation has been obtained, and an explanation of the formation of the soils has been offered.

Minerals varied in their powers of resistance to weathering in the order: oligoclase < sanidine < mica (muscovitic) = chlorite < expanding micas (vermiculite and montmorillonite) < kaolin < gibbsite. Rhyolitic volcanic glass was less stable than sanidine, and quartz was at least as stable as gibbsite. There was some evidence to suggest that the order of resistance to podzolisation was almost the reverse of the order above, but again quartz was very resistant. Secondary silica crystallised as cristobalite appeared to be the only mineral formed by podzolisation.

The sand fractions contained mainly primary minerals and indicated the relative stabilities of these minerals. The coarse and medium silts contained little other than quartz, but the fine silts contained both primary and secondary minerals and indicated the relative stabilities of the most stable primary minerals and the least stable secondary minerals. The clay fractions contained mostly secondary minerals, were the most susceptible to katamorphic changes, and indicated the relative stabilities of secondary minerals.

Koikoi sandy loam was formed under low to moderate weathering and low podzolisation on a steep slope. It contained nearly equal amounts of mica, kaolin and quartz in its clays. Maungarei clay loam was formed under strong weathering and low podzolisation and its clay was predominantly kaolin. Pukenamu silt loam was formed under strong weathering and moderate podzolisation and had definite eluvial and illuvial horizons. Its clays were predominantly kaolin and quartz with quartz concentrated in the eluvial horizons. Parahaki sandy loam was formed under strong weathering and strong podzolisation, and showed marked differentiation into eluvial and illuvial zones. Quartz was the dominant mineral in the eluvial horizons but gibbsite and kaolin were dominant in illuvial horizons. Anatase was concentrated in one illuvial horizon. The C horizon of this soil contained an unusual "stuffed" cristobalite with a high cation exchange capacity and high water holding capacity.

INTRODUCTION

In any study of soil formation, the nature of the parent rock must be considered because changes in the parent rock may cause important changes in the parent material and hence in the soils. Particularly is this so in a mineralogical study of soil formation, and it is advantageous to confine such an investigation to the soils developed from a single rock.

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In the present work four soils derived from rhyolite have been studied. The mineralogical compositions of the sand, silt and clay fractions have been determined and related to the amounts of weathering and podzolisation the soils have undergone.

REVIEW OF LITERATURE

McLaughlin (1955) examined some soils developed from an arkosic parent rock in New Zealand. He found kaolin and quartz in the clays of all the soils he examined. Feldspars and hydrous micas occurred in the clays of the less weathered soils. Costin, Hallsworth, and Woof (1952) examined a climosequence of soils developed on gneissic granite in Australia. The soils occurred in an altitudinal sequence from 3,000 ft to 7,300 ft, with vegetation ranging from schlerophyl forest in the lower levels to bunch grasses in the upper. Under these conditions an optimum weathering region occurred at about 5,500 to 6,000 ft. The clay minerals in the soils of this region were kaolin and gibbsite with some clay-biotite. Above and below this region the clays contained clay-biotite with 10–40% of kaolin and very little gibbsite.

Very few reports of the mineralogy of soils from rhyolite or of the weathering products of this rock appear in the literature. Cross (1896) observed that ryholite, under the influence of sulphurous waters, weathered to quartz and alunite, and that volcanic glass from the rock decomposed to opaline silica. Uchida (1944) records the formation by weathering of a bentonite from rhyolite in Japan.

Marel (1948), investigating the weathering changes in a rhyolitic tuff in Sumatra, observed that volcanic glass was fairly resistant to attack although the most weathered soils contained none of it. Of the heavy minerals present – hypersthene, amphibole, allanite, zircon, and ore minerals – only allanite and zircon accumulated in the sequence from tuff to strongly weathered soil. Hypersthene and amphibole progressively decreased while the ore and opaque minerals increased to a moderate weathering stage and then decreased sharply.

The mineralogy of soils and weathered materials from granites, adamellites and granodiorites have been studied to some extent. These rocks may be considered to be the coarse-grained equivalents of rhyolite. The clays of youthful soils developed from these rocks usually contain micas and expanding mica-like minerals. Thus Stephen (1952), in a detailed petrographic and X-ray study of a shallow, stony soil from granite, identified dioctahedral illite with some kaolinite and quartz. He considered the illite, some of which showed signs of expanding interlayers, to be derived from the feldspars in the rock, partly by weathering and partly by hydrothermal alteration. Microcline stayed fresh even into the topsoil.

Butler (1953) examined two young, immature soils developed from adamellite and granodiorite in the south of England. He found that the clays from both soils were similar, containing illite, interstratified illite-vermiculite, and kaolinite, with lesser amounts of residual quartz and feldspars. Butler considered the illite to be a sericitic product of the parent material. Humbert and Marshall (1943), using an electron microscope, also identified micas in the clays of a Grey Brown Podzolic soil from granite, although the predominant clay mineral was beidellite. Kaolinite and quartz also occurred, the latter increasing in amount from the unweathered rock to the surface. They also noted that microcline was more stable than other feldspars.

Under more aggressive conditions these rocks give rise to kaolinic clays. Cady (1951) found predominantly kaolinitic minerals at all depths down to 15 ft in a soil developed from granodiorite in the Piedmont region of North Carolina. There was much more quartz, some of which was secondary, in the surface horizons of this soil than there was at depth. Cady considered that the first weathering products of the rocks were gibbsite, chlorite and allophane, and that these were later changed to kaolinite, secondary quartz and iron oxides. Hosking (1940) found kaolinite in the subsoils of a number of Australian soils from granite.

Under the weathering conditions which lead to the formation of laterite, granite gives rise to clays containing kaolinic minerals and quartz. This has been noted by Harrison (1933), Lapparent (1939), Humbert (1948), Carroll and Jones (1947), and no doubt many others. Allen (1952) examined a bauxite deposit derived from granite in Georgia and observed that kaolinite was formed from some of the primary minerals and later altered to the gibbsite which made up the main ore body. Microcline and albite remained fresh and unaltered throughout the rock to bauxite sequence.

Alexander, Hendricks, and Nelson (1939) obtained semiquantitative estimates of the mineral colloids in soils developed from granitic parent materials in the United States. They examined Desert soils, Grey Brown Podzolic soils, and one Red Podzolic soil, so that their results serve somewhat as a summary of the foregoing literature. In the Desert soils the clays contained 20% kaolinite and 70% hydrous mica-montmorillonite. In the Grey Brown Podzolic soils the clays contained 50% kaolinite and 30–40% hydrous mica. In the Red Podzolic soil the clays contained 80% kaolinite and no hydrous mica.

DESCRIPTION OF THE RHYOLITES

The rhyolites chosen for the present study were required to conform as closely as possible to the following criteria:

(1) They must have similar chemical compositions;
(2) They must have similar mineralogical compositions;
(3) They should show no evidence of hydrothermal alteration;
(4) They should be similar in texture and structure; and
(5) They should occur in localities where there is little likelihood of contamination by wind-blown or alluvial materials.

Three rhyolites which conformed to the above criteria were finally selected. They were the rhyolites of Mt. Somers, described by Speight (1938); of Mangawai, described by Ferrar (1934); and of Putahi Hill, described by Bell and Clarke (1909) and Ferrar (1934).

The Mangawai and Putahi Hill rhyolites were slightly more dacitic than the Mt. Somers rhyolite, and contained more Na_2O and less K_2O. One sample of the Mangawai rock contained primary muscovite in place of sanidine. Otherwise the rocks were similar in chemical and mineralogical composition. No alteration products of feldspar were identified in the sections, nor in X-ray patterns of the crushed rock, but a small amount of chlorite, apparently produced from the volcanic glass, was identified petrographically in some of the sections. It is unlikely that alteration products will be completely absent from the rocks, and some of the secondary minerals present in the soils will have been produced by katamorphic processes other than weathering. Every effort has been made to keep these amounts to a minimum.

The greatest differences between the rhyolites were in their textures. Although all three had porphyritic textures, the phenocrysts ranged in size from 1 mm to 0·1 mm, and the groundmass varied from wholly crystalline to non-crystalline. It is considered that these textural differences may cause differences in the rates of weathering, but the products of weathering in all but the earliest stages should be the same regardless of the rock texture.

The Koikoi soil occurs in a region where greywacke loess mantles are found. It is believed that little loess has been deposited on this site at 2,500 ft altitude and on a 28° slope and that any that may have been so deposited has been removed by erosion. There is no sign of it in the present soil. It is possible, however, that some of the micaceous mineral colloids in this soil were derived from loess.

Petrographic description of rock samples from beneath or beside the soils sampled, chemical analyses, and detailed discussions of the extent to which each rock conformed to the set of criteria may be found in the thesis by Swindale (1955).

DESCRIPTION OF THE SOILS

Four sedentary soils were collected from the three rhyolite flows. Their main soil-forming factors and classification are summarised in table 1.

Koikoi Fine Sandy Loam

Locality: West bank of Woolshed Creek, Mt. Somers, S81 : 832469.
Slope: 28°. *Altitude:* 2,500 ft.

Profile Description:

A	6 in.	pale brown (2.5Y-10YR 5/3) fine sandy loam; loose; weakly developed fine crumb and fine to medium nutty structure; much root penetrated; merging to
C	8 in.	yellowish brown (2.5Y-10YR 5/4) fine sandy loam; friable; weakly developed fine blocky structure; many rock fragments and boulders through the profile
	on	hard rhyolite.

Classification: Weakly weathered, yellow-brown earth.
Parent Rock: Mt. Somers rhyolite.

At present the area is under sub-alpine grasses and scrub and is used for extensive sheep-farming. Past overstocking and frequent burning have led to severe sheet erosion, and 50% of the area is bare of plant cover.

Maungarei Clay Loam

Locality: Near Maungaturoto junction on main highway, N28 : 946521.
Slope: 12°.

Profile Description:

A_1	5 in.	dark brown (10YR 3/3) clay loam; friable; weakly developed, fine nutty structure; some worm-cast granules and numerous grass roots,
A_3	7 in.	light yellowish brown (10YR 6/4) clay loam; friable to firm; common, fine, faint light grey and reddish brown mottlings; moderately developed medium nutty structure with some worm cast granules, fairly numerous grass roots,
BC	7 in.	brownish yellow (10YR 6/6) clay; firm; few, fine faint grey and reddish brown mottlings; moderately to strongly developed coarse nutty structure; fairly numerous grass roots,
C	12 in.	brownish yellow (10YR 6/6) clay, firm to very firm; strongly developed, coarse nutty structure; compact; few fine roots.
	on	brownish yellow very firm clay.

Classification: Strongly weathered, yellow-brown earth.

Parent Rock: Mangawai rhyolite.

Developed under evergreen broadleaf forest. At present area is under broadleaf trees (*Beilschmiedia taraire, Vitex luscens,* etc.) and grasses.

Pukenamu Silt Loam

Locality: 2½ miles north of Hakaru on Browns Road, N28 : 105503.

Slope: < 5°.

Profile Description:

A_2	6 in.	dark grey (10YR 4/1) silt loam; friable; massive structure; numerous grass roots,
B_1	4 in.	grey (10YR 5/1) silty clay; firm, common, fine to medium, distinct reddish brown mottlings, massive structure with tendency to weakly developing nutty structure; fairly numerous scrub roots; sharp boundary,
B_2	22 in.	pale grey (2.5Y 7/2) clay; very firm; many, medium to coarse, prominent, brownish yellow and reddish yellow mottlings; strongly developed coarse nutty structure, massive in places; humus staining on blocks.
C_1	3 ft	pale grey (2.5Y 7/2) clay; very firm; common, medium to fine, distinct yellowish brown mottlings; strongly developed coarse nutty structure; humus staining on faces of blocks.
	on	white clay.

Classification: Moderately podzolised, strongly weathered, yellow-brown earth.

Parent Rock: Mangawai rhyolite.

Developed under kauri (*Agathis australis*) forest. At present under manuka scrub.

Parahaki Sandy Loam

Locality: Te Pua Rd, just south of Putahi Hill, N15 : 325389.

Slope: < 5°.

Profile Description:

A_1	11 in.	dark grey (10YR 4/1); sandy loam; friable; massive and single-grained structure; fairly numerous scrub roots,
A_2	4½ in.	pale grey (2.5Y 7/2); (white, humus stained) sand; structureless; strongly cemented,
A_3–B_{21}	3 in.	black (5YR 2/1); structureless loam . . . humus layer,
B_{22}	5 in.	10 mm of dark reddish brown (2.5YR 3/4) indurated material; transition to strong brown (7.5YR 5/6) clay; gritty, sticky; weakly cemented and in places coarsely laminated,
B_{23}	12 in.	pale yellow (2.5Y 7/4) clay; firm; fine, faint mottlings; massive structure showing some fragmental structure when disturbed; sticky,
C_1	on	greenish brown, strongly weathered clay.

Classification: Strongly podzolised, strongly weathered, yellow-brown earth.
Parent Rock: Putahi Hill rhyolite.
Developed under kauri forest. At present under low scrub.

EXPERIMENTAL METHODS

The samples taken (see table 1) were air-dried and crushed to pass a 2-mm sieve. Organic matter was removed by treatment with H_2O_2, and extractable iron oxides were removed and determined by the method of Aguilera and Jackson (1953). The samples were then separated into size fractions by the method of Jackson, Whittig, and Pennington (1950), in which dispersion is aided by boiling the sample for 5–10 minutes in 2% Na_2CO_3. The fractions, coarse sand ($> 500\ \mu$); medium sand ($500–250\mu$); fine sand ($250–105\mu$); very fine sand ($105–50\mu$); coarse silt ($50–20\mu$); medium silt ($20–5\mu$); fine silt ($5–2\ \mu$); coarse clay ($2–0{\cdot}2\ \mu$); medium clay ($0{\cdot}2–0{\cdot}08\ \mu$); and fine clay ($< 0{\cdot}08\ \mu$), were obtained by a combination of sieving and sedimentation techniques.

TABLE 1—Environment and Classification of Soils and Horizons Analysed

Soil Type	Annual Rainfall	Average Temperature	Soil-forming Vegetation	Slope	Classification	Horizon Analysed	Depth
	in.	°F		degrees			in.
Koikoi fine sandy loam	40	47·5	Beech forest and grasses	28	Weakly weathered yellow-brown earth	A C	0–6 6–14
Maungarei clay loam	60·4	57	Broadleaf forest	12	Strongly weathered yellow-brown earth	A_1 A_3 C_1	0–5 5–12 21–31
Pukenamu silt loam	47·7	57	Kauri forest	< 5	Strongly weathered moderately podzolised yellow-brown earth	A_2 B_1 B_2 C_1	0–6 6–10 10–29 34–42
Parahaki sandy loam	61·8	57·5	Kauri forest	< 5	Strongly weathered strongly podzolised yellow-brown earth	A_1 A_2 $A_3–B_{21}$ B_{22} B_{23} C_1	0–11 11–15½ 15½–18½ 19–23 29–36 41–49

The sand fractions were examined petrogaphically, and counts of mineral grains in the very fine sand fractions were made. The coarse and medium silt fractions were examined petrographically, by means of X-ray diffraction, and their cation exchange capacities were determined. The fine silt fractions were examined more fully, and were subjected to the same methods of analysis as those used for the examination of the clay fractions.

The clay in each horizon was separated into three fractions to facilitate the identification of all the minerals contained therein. Semiquantitative estimates of the mineralogical compositions of these fractions were obtained from various X-ray diffraction techniques, differential thermal analysis, heating - weight-loss studies, determinations of cation exchange capacities, and determinations of specific surface areas. The compositions so obtained were then refined by adjusting the figures so that the calculated chemical composition of each fraction agreed with a chemical composition obtained experimentally. Standard chemical formulae for the minerals were obtained from various texts on clay mineralogy and used in this refining procedure.

Unoriented powder specimens and basally oriented specimens (oriented to enhance the intensities of the X-ray reflections from the basal planes) were used in the X-ray identifications. To help identify the various layer silicates, X-ray diffraction patterns were obtained from specimens which had been: (a) Mg-saturated and glycerol solvated; (b) K-saturated; and (c) K-saturated and heated to 350, 500, and 600°C.

In heating - weight-loss studies, the weight lost by samples in the intervals: 110–350°C, 350–500°C, and 500–900°C were obtained as percentages of the weight at 110°C. Weight losses in these temperature intervals give approximate estimates of the respective percentages of gibbsite, kaolin, and 2 : 1 and 2 : 2 layer silicates in the samples. Cation exchange capacities were determined by the method of Swindale and Fieldes (1952). Specific surface areas were obtained by a method which employed glycerol absorption upon unignited samples from an anhydrous medium. The glycerol absorbed by the sample was determined volumetrically. Chemical analyses were carried out by the method of Corey and Jackson (1953) with slight modifications to allow the determinations of Na_2O, K_2O, MgO, and CaO to be made by flame photometry.

MINERALOGICAL NOMENCLATURE

It has been deemed advisable to describe the mineral names used in this paper.

Quartz, sanidine, oligoclase, gibbsite, and anatase have their usual meanings. Cristobalite includes both the pure silica mineral and the "stuffed" structure of Buerger (1954).

Mica: This name was used for a 2 : 1 layer silicate with a 001 spacing of approximately 10 Å. This spacing, as exhibited in X-ray diffraction patterns, was unaffected by heating the sample to 500°C, by saturating the sample with various cations, or by glycerol-solvating the sample. In the rocks and soils investigated in this work, where biotite was found only as occasional shreds, the mica was dioctahedral.

Chlorite: This name was used for a 2 : 2 layer silicate with a 001 spacing of approximately 14 Å. This spacing was unaffected by heating the sample to 500°C, by saturating the sample with various cations, or by glycerol solvation.

Vermiculite: This name was used for a 2 : 1 layer silicate with a 001 spacing which could vary between 14 Å and 9·5 Å according to the pretreatment that the sample received. A Mg-saturated, glycerol-solvated sample had a 14 Å spacing; a K-saturated sample which had been heated to 500°C had a spacing between 9·5 and 10 Å. No distinctions in nomenclature were made between dioctahedral minerals and trioctahedral minerals with these properties.

Montmorillonite: A 2 : 1 layer silicate with a 001 spacing which could vary between 18 Å and 9·5 Å according to the pretreatment that the sample received. A Mg-saturated, glycerol-solvated sample had a spacing of approximately 18 Å; a K-saturated sample which had been heated to 500°C had a spacing between 9·5 and 10 Å. No nomenclatural distinction was made between dioctahedral and trioctahedral montmorillonites.

Kaolin: This name was used for a 1 : 1 layer silicate with a 001 spacing which could vary between 11 and 7 Å according to the pretreatment that the sample received, although for the most part the variation was restricted to the range 7·5 Å to 7 Å. A K-saturated sample gave a spacing between 7·5 and 7 Å which disappeared completely from X-ray diffraction patterns when the sample was heated to 500°C. Most of the samples also gave a 7·5 to 7 Å spacing when they were saturated with Mg and glycerol-solvated, but some of them contained sufficient hydrated halloysite to give spacings between 8·5 and 11 Å after this pretreatment.

A distinction between kaolinite and halloysite has been made in the results sections when it appears relevant. When such distinction is made the names have been used in the manner suggested by Brindley (1951).

Interstratified Mixtures: These were recognised frequently in the samples investigated. They were identified as interstratified mixtures of the minerals defined above. For example, a sample which gave a 12 Å spacing on Mg-saturation and glycerol-solvation, and a 10 Å spacing on K-saturation and heating to 500°C was called an interstratified mixture of mica and montmorillonite or mica and vermiculite. The choice between these two would depend upon other evidence.

It should be obvious from the preceding discussion that identifications of minerals were made primarily by X-ray diffraction methods. These methods, however, are only semiquantitative at best, and are not even as good as that when mixtures such as chlorite-kaolin or randomly interstratified mica-vermiculite-montmorillonite occur. Consequently a combination of methods was used to obtain the mineralogical compositions shown in the tables of this paper.

RESULTS

Koikoi Fine Sandy Loam

The main difference between the two horizons of this shallow stony soil was the appearance of organic matter in the topsoil, and the profile was that of an A-C soil. There was little difference between the particle size distributions or between the extractable iron oxide contents of the two horizons (see table 2).

Mineralogy of the Sand and Silt Fractions: Primary quartz, feldspars, and volcanic glass made up the greater part of the sand fractions. Sanidine and oligoclase were both found in these fractions, but the relative amounts of the two were different in the two horizons. Thus the oligoclase/sanidine ratio for the very fine sand fraction of the C horizon was 3·36, while the same ratio for the A horizon was only 1·55.

Quartz and feldspars were also the most numerous minerals in the coarse and medium silts, but mica, chlorite, vermiculite, and montmorillonite all occurred in small quantities. There appeared to be slightly more sanidine than oligoclase in these fractions but the differences were not great.

TABLE 2—Particle Size Distributions and Extractable Iron Oxide Percentages of the Soils

Horizon	Coarse Sand $> 500\mu$ %	Medium Sand $500–250\mu$ %	Fine Sand $250–105\mu$ %	Very fine Sand $105–50\mu$ %	Coarse Silt $50–20\mu$ %	Medium Silt $20–5\mu$ %	Fine Silt $5–2\mu$ %	Coarse Clay $2–0{\cdot}2\mu$ %	Medium Clay $0{\cdot}2–0{\cdot}08\mu$ %	Fine Clay $< 0{\cdot}08\mu$ %	Extractable Fe_2O_3 %
					Koikoi Fine Sandy Loam						
A	7·85	4·75	7·45	10·25	22·95	21·15	8·00	12·05	4·00	1·40	1·10
	10·30	5·65	8·15	9·60	21·80	19·90	7·75	11·45	3·85	1·70	1·07
					Maungarei Clay Loam						
A_1	0·00	0·25	0·80	10·60	47·75	16·40	2·45	9·35	10·95	1·45	0·41
A_3	0·00	0·00	0·80	14·20	41·40	15·70	2·70	10·60	13·70	1·40	0·63
C_1	0·00	0·25	1·90	11·30	33·20	20·80	4·80	15·95	11·55	2·10	1·23
					Pukenamu Silt Loam						
A_2	0·18	0·39	1·70	2·30	22·10	44·50	6·10	18·20	2·55	2·10	0·07
B_1	0·76	0·32	1·15	2·09	18·33	45·44	6·05	17·29	4·32	4·25	0·28
B_2	0·00	0·00	0·50	1·15	18·50	31·00	6·20	24·55	11·50	6·75	0·77
C_1	0·00	0·00	0·00	0·15	8·25	50·90	3·85	21·40	13·95	1·65	0·93
					Parahaki Sandy Loam						
A_1	12·60	16·85	12·75	4·65	8·00	31·00	8·10	5·85	0·15	0·00	0·01
A_2	17·77	18·62	13·01	4·33	6·70	26·59	7·11	5·79	0·06	0·00	0·02
A_3–B_{21}	19·00	19·50	12·80	3·20	5·60	24·00	6·00	8·35	1·30	0·35	0·53
B_{22}	5·35	14·95	12·40	5·65	10·70	22·80	10·10	11·85	4·90	1·00	5·35
B_{23}	2·10	5·75	3·00	2·20	6·26	29·10	17·15	16·55	17·00	0·95	0·78
C_1	4·60	14·00	8·20	4·20	15·20	26·45	18·55	8·45	0·30	0·00	0·73

Analytical data for the fine silt fraction (5–2 microns) of both horizons are shown in table 3. The data for the two horizons are very similar, and show a big reduction in SiO_2 from the parent rhyolite which contained 77% SiO_2. X-ray diffraction patterns for the A horizon, obtained with basally oriented specimens, are represented in fig. 1a. The patterns indicate the mineralogical complexity of these fractions.

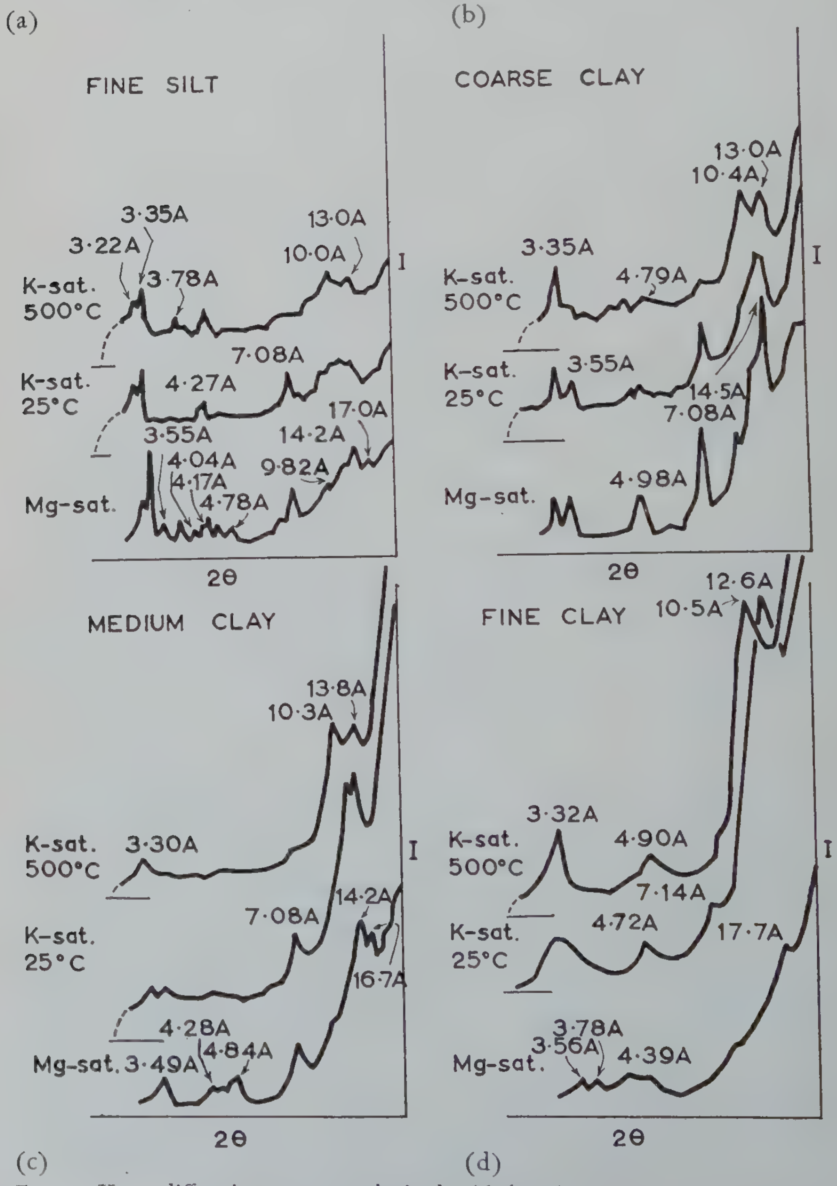

FIG. 1—X-ray diffraction patterns, obtained with basally oriented specimens, for the fine silt and clay fractions of Koikoi A horizon.

The mineralogical compositions are shown in table 4. Sanidine, oligoclase, and quartz all occurred in amounts between 20 and 30%. The oligoclase/sanidine ratio, unlike that of the very fine sand fraction varied little between the A and C horizons. Numerous layer silicates were found in these fractions, the total amounts of these minerals being approximately 30%. A small fraction of these could have been derived from contaminating loess, but most of them must be considered secondary minerals. There was more vermiculite and montmorillonite and less mica and chlorite in the A horizon than there was in the C.

Mineralogy of the Clay Fractions: Analytical data for the clay fractions are shown in table 3, and mineralogical compositions in table 4. X-ray diffraction patterns for basally oriented samples from the A horizon are represented in figs. 1b, c, d.

The most obvious difference between the fine silt and coarse clay fractions was the near absence of feldspars in the latter. This difference can be seen by comparing figs. 1a and 1b. These figures also reveal the larger amounts of layer silicates in the coarse clays. Fig. 1b has intense diffraction maxima at 14 Å in the patterns from glycerol-solvated samples. In the patterns from K-saturated, heated samples there are two maxima at 10 and 13 Å. From these patterns it is inferred that vermiculite, which contributed to the 14 Å spacing in the glycerol-solvated sample and collapsed to 10 Å on heating at 500°C, was present in significant amounts. The diffraction maximum at 13 Å in the heated sample and the low "shoulder" at 13 Å in the glycerol-solvated sample indicate that chlorite was present in interstratified mixtures with mica and vermiculite.

There was 19% kaolin in the coarse clay of the A horizon and 24% in the coarse clay of the C. These relatively large amounts of kaolin indicate the importance of the bulk chemical composition of the rhyolite to the nature of the katamorphic products. Oxides of silicon and aluminium, produced by weathering of primary minerals, can recombine to form secondary minerals containing iron, magnesium, calcium, and potassium, only when these elements are present. In their absence the only secondary minerals possible are oxides and aluminium silicates. It is not surprising, therefore, to find gibbsite and kaolin as early weathering products of a rock such as rhyolite which contains more than 85% of silica plus alumina and relatively little iron, magnesium, calcium, and potassium.

The medium clay fractions of both horizons contained the same materials as the coarse clay fractions, and montmorillonite and gibbsite as well. The amounts of quartz and mica were less than in the coarse clay and the amounts of expanding layer silicates greater. There was more quartz in the medium clay of the A horizon than there was in the medium clay of the C horizon.

There was insufficient fine clay in either horizon to allow chemical analyses to be made, and the estimates of mineral quantities were less accurate because of this. The fine clays appeared similar, containing 60% montmorillonite, with some kaolin, gibbsite, vermiculite, and mica. The montmorillonite in these fractions was clearly dioctahedral. X-ray diffraction patterns, obtained with basally oriented samples heated to 500°C, showed strong 001 and 003 maxima and moderate 002 maxima. The latter would have been very weak or absent if the montmorillonite had been trioctahedral.

TABLE 3—Cation Exchange Capacities, Specific Surface Data, Data on Heating Weight Loss, and Chemical Analyses of the Fine Silt and Clay Fractions of Koikoi Fine Sandy Loam and Maungarei Clay Loam

Horizon	Size Fraction μ	C.E.C. M.E. %	Specific Surface*	Heating-weight Loss			Chemical								
				110–350° %	350–500° %	500–900° %	SiO_2 %	Al_2O_3 %	Fe_2O_3 %	FeO %	MgO %	CaO %	Na_2O %	K_2O %	TiO_2 %
							Koikoi Fine Sandy Loam								
A	5–2	18.2	2.6	1.73	1.32	1.07	66.62	18.39	1.34	1.35	1.12	0.54	2.76	4.10	0.29
	2–0.2	48.0	5.2	2.78	3.26	2.25	53.55	23.71	3.71	2.17	2.52	0.34	1.02	3.83	0.94
	0.2–0.08	85.5	10.1	7.20	4.41	2.85	45.09	29.24	2.98	1.86	2.04	0.24	1.37	2.24	0.97
	< 0.08		18.0	7.7	6.1	5.9									
C	5–2	19.1	2.1	1.43	1.27	1.43	67.13	16.57	0.85	1.54	1.00	0.30	3.21	4.33	0.34
	2–0.2	43.7	6.2	3.08	4.04	2.29	51.62	23.44	3.91	1.98	2.51	0.50	1.08	3.91	0.61
	0.2–0.08		10.0	6.47	4.57	2.39	43.89	29.97	4.48	2.02	2.44	0.22	0.42	2.71	1.42
	< 0.08		17.7	9.4	4.4	5.9									
							Maungarei Clay Loam								
A1	5–2	8.4	5.4	2.4	1.7	0.2	59.81	29.02	1.28	1.05	0.56	0.00	0.18	0.31	1.43
	2–0.2	27.2	5.6	3.12	7.70	2.11	48.68	37.90	1.43	0.59	0.43	0.00	0.25	0.00	0.38
	0.2–0.08	33.5	5.9												
	< 0.08		12.0	1.5	9.7	5.5									
A3	5–2	9.5	3.2	2.7	0.9	0.7	54.80	29.54	1.62	0.70	0.40	0.00	0.51	0.21	1.30
	2–0.2	26.9	6.2	2.22	7.88	1.35	46.38	34.22	1.59	0.63	0.21	0.00	0.00	0.00	0.42
	0.2–0.08	33.8	6.8	1.98	10.60	3.32									
	< 0.08		12.1	1.4	9.9	4.8									
C1	5–2	33.5	7.9	1.93	7.22	1.58	54.16	32.08	1.12	1.00	0.81	0.00	0.87	0.11	0.37
	2–0.2	43.8	6.0	1.83	9.34	2.91	48.32	36.18	1.21	0.82	0.52	0.00	0.27	0.00	0.45
	0.2–0.08	35.7	7.2	2.25	10.85	2.25	47.58	36.75	1.81	0.82	0.14	0.00	0.00	0.00	0.44
	< 0.08		12.0	4.3	9.3	4.5									

*The units for specific surface are mg glycerol per 50 mg sample.

TABLE 4—Mineralogical Compositions of the Fine Silt and Clay Fractions of Koikoi Fine Sandy Loam and Maungarei Clay Loam

Horizon	Size Fraction μ	Amount of Fraction in Horizon %	Sd %	Ol %	Qr %	Cb %	Mc %	Ch %	Vr %	Mt %	Kl %	Gb %	An %
					Koikoi Fine Sandy Loam								
A	5–2	8.00	21	22	31		5	5	4	4	8	2	
	2–0.2	12.05			24		33	9	12		19	2	1
	0.2–0.08	4.00			14		19	8	9	10	26	10	2
	< 0.08	1.40					5		5	60	20	10	
C	5–2	7.75	19	24	30		10	6	2		8		
	2–0.2	11.45	5		19		27	9	13		24		1
	0.2–0.08	3.85			7		23	9	14	14	24	6	1
	< 0.08	1.70					5		5	60	20	10	
					Maungarei Clay Loam								
A_1	5–2	2.45			65	5	2	3	3		20	2	
	2–0.2	9.35			28		3	4	3	1	56	5	1
	0.2–0.08	10.95			5			2	4		89		
	< 0.08	1.45						8	3	5	80	4	
A_3	5–2	2.70			60	5	2	3	3		25	2	
	2–0.2	10.60			21		2	1	4	4	63	3	1
	0.2–0.08	13.70			5			2	4		87		
	< 0.08	1.40						7	7	7	75	4	
C_1	5–2	4.80			16	5	1	4	3		70		
	2–0.2	15.95			5			3	3		88		1
	0.2–0.08	11.55			5				5		90		
	< 0.08	2.10			2		2	5	7	8	66	10	

Sd = sanidine; Ol = oligoclase; Qr = quartz; Cb = cristobalite;
Mc = mica; Ch = chlorite; Vr = vermiculite; Mt = montmorillonite.
Kl = kaolin; Gb = gibbsite; An = anatase.

Maungarei Clay Loam

This soil, a strongly weathered yellow-brown earth, had no recognisable A_2 horizon. Three horizons, designated A_1, A_3, and C_1, were sampled and investigated. The clay contents and the extractable iron oxide contents of these three horizons increased with depth (see table 2).

Mineralogy of the Sand and Silt Fractions: Quartz, oligoclase, and a small quantity of potash feldspar were the only minerals which were found in the sand fractions of this soil. As in the Koikoi soil, the oligoclase/sanidine ratio decreased towards the surface.

Quartz was the predominant mineral in the coarse and medium silt fractions. Cristobalite and the feldspars occurred in very small amounts. Layer silicates with basal spacings of 14, 10, and 7 Å were present, and became more abundant with depth.

Quartz was also the predominant mineral in the fine silt fractions of the A_1 and A_3 horizons, but in the C_1 horizon a change occurred and kaolin became predominant. Cation exchange capacities and data on heating weight loss (table 3) indicate this change, but the X-ray diffraction patterns obtained with basally oriented specimens (figs. 2a, b, d) do not show the expected increase in kaolin spacings for the C horizon because the kaolin was halloysite which cannot be oriented to enhance reflections from basal spacings.

Mineralogy of the Clay Fractions: Analytical data and mineralogical compositions of the clay fractions of the three horizons sampled are shown in table 3 and table 4 respectively. Some typical X-ray diffraction patterns obtained with basally oriented specimens are presented in fig. 2 and fig. 3.

Kaolin was the predominant mineral in the coarse clay fractions of the three horizons. Quartz was the only other mineral which occurred to any extent. The amount of quartz decreased sharply between the A_3 and C_1 horizons as the amount of kaolin increased. Mica, chlorite, and the expanding layer silicates occurred in small amounts in all horizons.

The medium and fine clays were mainly composed of halloysitic kaolin. There appeared to be more expanding layer silicates and more chlorite in the fine clay than in the coarser fractions, and in the C_1 horizon there appeared to be gibbsite which was estimated at 10%.

Pukenamu Silt Loam

The Pukenamu soil was developed under a coniferous (kauri) forest under an aggressive mor litter and is moderately podzolised although strongly weathered. The conifers of northern New Zealand, and the kauri (*Agathis australis*) in particular, are apparently unique in their ability to produce strongly podzolised soils in regions of strong weathering. The Pukenamu soil is moderately podzolised because the kauri forest beneath which it formed was of no great age. The profile sampled had an A_2 horizon which was 6 in. deep, and which was darker than usual owing to the assimilation of organic matter from the present scrub vegetation. It contained 23% of clay, mostly as coarse clay, and 73% of silt. There was very little sand in this horizon or in any of the horizons sampled. The absence of sand was partly due to weathering and partly due to the finely crystalline nature of the parent rock. The B_2 horizon was only moderately developed, and contained 5% more clay than the C horizon below. The soil did not show any accumulation of extractable iron oxides, but there was a sharp increase in these from the B_1 to the B_2 horizons (see table 2).

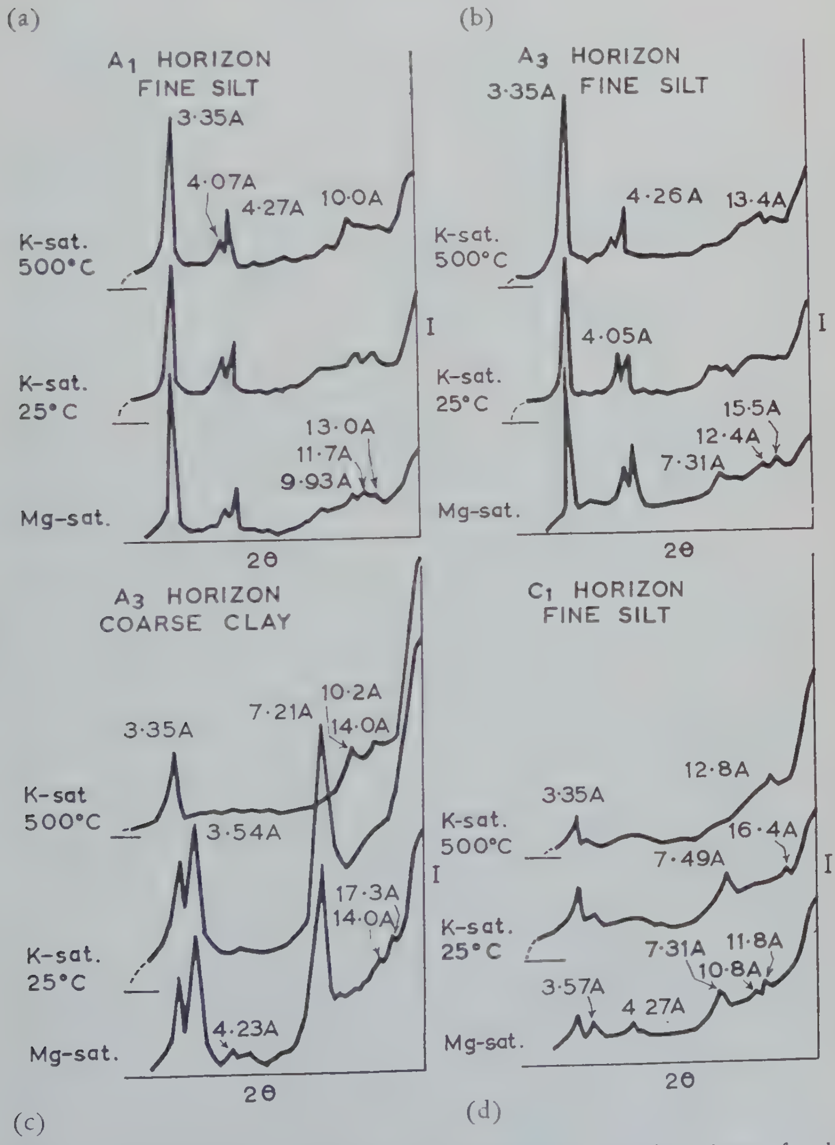

FIG. 2—X-ray diffraction patterns, obtained with basally oriented specimens, for the fine silt fractions of the Maungarei A_1, the fine silt and coarse clay fractions of the A_3 horizon, and the fine silt fraction of the C_1 horizon.

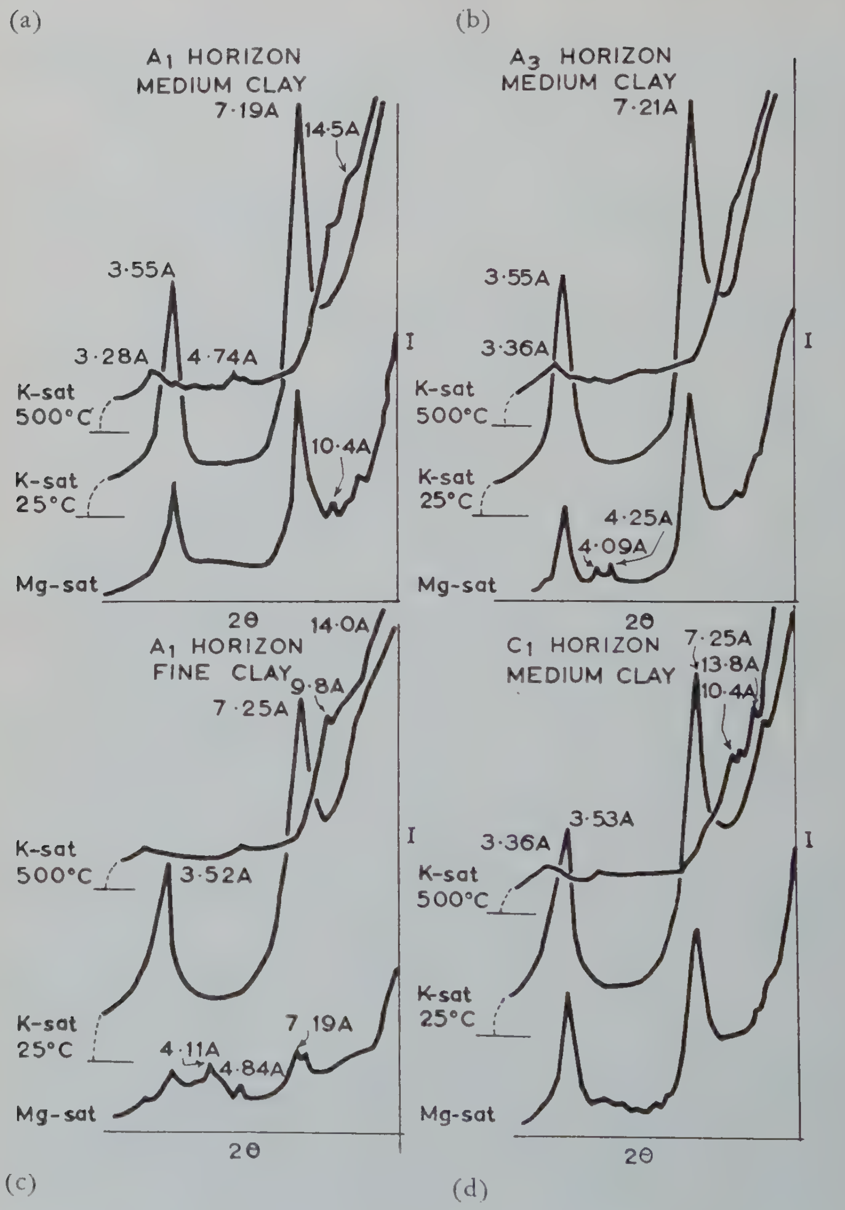

FIG. 3—X-ray diffraction patterns, obtained with basally oriented specimens, for the medium and fine clay fractions of the Maungarei A_1 horizon, the medium clay fraction of the A_3 horizon, and the medium clay fraction of the C_1 horizon.

Mineralogy of the Sand and Silt Fractions: Quartz was the dominant mineral in the sand fractions, and in the coarse and medium silt fractions.

Volcanic glass, cristobalite, rock aggregates, and small quantities of layer silicates were also present. The percentages of minerals other than quartz increased with increasing depth and with decreasing particle size.

Analytical data for the fine silt fractions are recorded in table 5 and mineralogical compositions in table 6. X-ray diffraction patterns obtained with basally oriented samples for the B_2 and C_1 horizons are shown in figs. 4a, and b. Quartz comprises practically all the fine silt of the A_2 and B_1 horizons, and a large proportion of the fine silt of the B_2 horizon as well. An increase in halloysitic kaolin occurs from 10% in the B_2 horizon to 48% in the C_1 horizon in which kaolin is the predominant mineral. Small percentages of the other layer silicates occur in the A_2, B_1, and B_2 horizons, and vermiculite plus montmorillonite comprise 9% of the fine silt of the C_1 horizon.

Mineralogy of the Clay Fractions: Analytical data and mineralogical compositions of the clay fractions are shown in tables 5 and 6 respectively. Some typical X-ray diffraction patterns obtained with basally oriented samples are reproduced in figs. 4c and d, and fig. 5.

Quartz was the predominant mineral in the coarse clay fractions of the A_2, B_1, and B_2 horizons (see table 6), but there was 22% kaolin in the B_2 horizon, and kaolin became predominant in the coarse clay of the C_1 horizon. Small amounts of 2 : 1 and 2 : 2 layer silicates occurred in all horizons, and there was a slight tendency for the expanding minerals to increase in amount with depth.

There was much less quartz in the medium clay fractions, and kaolin was the predominant mineral in all horizons. The percentages of both vermiculite and montmorillonite were greater than in the coarser fractions, and there were significant amounts of mica and chlorite in most horizons. Small quantities of gibbsite occurred in the lower horizons.

The X-ray diffraction patterns for the glycerol-solvated specimens did not show diffraction maxima corresponding to the 001 spacings of vermiculite and montmorillonite (see figs. 5a and b). These patterns did show much modification of the background scatter, however, and it seems reasonable to infer that the two minerals are interstratified together. Inclusion of mica in the interstratified mixture, variations in the percentages of the two expanding minerals, and differences in the structure factors of individual layers could all contribute to the absence of definite maxima. Sharp maxima at 10 Å were obtained in the heated, K-saturated samples.

Only in the B_1 and B_2 horizons was there sufficient fine clay to allow full sets of analyses to be made. The mineral estimates for the other two horizons were obtained from comparisons of X-ray diffraction and other data with those for the more accurate estimates in the two B horizons. X-ray diffraction patterns for the fine clay of the B_1 and C_1 horizons are shown in figs 5c and d. The fine clays of all four horizons had similar compositions. Montmorillonite, kaolin, and vermiculite were the most abundant minerals, with smaller amounts of mica and chlorite. No quartz was identified in any of these fractions although small amounts of it were calculated from the chemical analyses for the two B horizons.

TABLE 5—Cation Exchange Capacities, Specific Surface Data, Data on Heating Weight Loss, and Chemical Analyses for the Fine Silt and Clay Fractions of Pukenamu Silt Loam

Horizon	Size Fraction μ	C.E.C. M.E. %	Specific Surface *	Heating-weight Loss			Chemical								
				110–350° %	350–500° %	500–900° %	SiO_2 %	Al_2O_3 %	Fe_2O_3 %	FeO %	MgO %	CaO %	Na_2O %	K_2O %	TiO_2 %
A_2	5–2	3·7	1·7	1·19	0·00	0·00	97·28	1·88	0·37		0·00	0·17	0·16	0·21	0·98
	2–0·2	8·4	1·5	1·21	1·26	0·65	91·43	3·98	0·53		0·24	0·00	0·10	0·09	1·16
	0·2–0·08		11·9	5·5	7·2	5·0									
	< 0·08		16·8	1·6	6·5	7·3									
B_1	5–2	7·8	1·2	0·29	0·58	0·10	97·38	1·46	0·36		0·00	0·03	0·14	0·07	0·99
	2–0·2	6·6	2·2	0·67	1·62	0·49	92·10	3·60	0·50		0·38	0·27	0·83	0·26	1·03
	0·2–0·08		9·5	2·19	8·24	2·05	60·52	21·32	2·82		1·04	0·05	0·18	0·76	0·75
	< 0·08		14·8	2·94	6·73	4·69	46·73	26·73	4·00	0·67	1·89	0·14	0·08	0·69	0·22
B_2	5–2	7·8	2·7	0·25	1·60	0·67	91·80	4·26	0·63		0·10	0·01	0·11	0·06	1·02
	2–0·2	19·0	4·5	1·15	3·66	0·90	80·92	10·29	0·85		0·28	0·00	0·20	0·32	0·88
	0·2–0·08		9·9	3·08	8·11	3·06	49·73	33·27	2·29	0·45	1·07	0·37	0·44	0·43	0·41
	< 0·08		15·4	5·64	8·55		50·82	30·19	3·26	0·64	1·54	0·00	0·08	0·38	0·21
C_1	5–2	25·1	5·8	0·30	8·51	0·91	65·64	21·27	1·54		0·28	0·00	0·29	0·06	0·60
	2–0·2	25·0	5·9	0·92	9·36	1·96	54·70	29·79	1·63		0·41	0·00	0·39	0·07	0·53
	0·2–0·08		10·0	2·16	8·18	2·32		34·87	1·82	0·44	0·56	0·06	0·24	0·29	0·40
	< 0·08		16·6	2·1	10·8	4·0									

*The units for specific surface are mg glycerol per 50 mg sample.

TABLE 6—Mineralogical Compositions of the Fine Silt and Clay Fractions of Pukenamu Silt Loam

Horizon	Size Fraction μ	Amount of Fraction in Horizon %	Qr %	Cb %	Mc %	Ch %	Vi %	Mt %	Kl %	Gb %	An %
A_2	5–2	6.10	90	3	2	2	2		3		1
	2–0.2	18.20	86		1		2	2	7		1
	0.2–0.08	2.55	15				15	25	45		
	< 0.08	2.10			5	5	20	35	30		
B_1	5–2	6.05	94		1	2			3		1
	2–0.2	17.29	98		2		2	2	5		1
	0.2–0.08	4.32	26	5	7	3	12	14	30	2	1
	< 0.08	4.25	2		6	3	14	40	28	2	
B_2	5–2	6.20	84	2	1	1	1		10		1
	2–0.2	24.55	68		3		3	2	22		
	0.2–0.08	11.50	8		4	2	8	20	53	5	1
	< 0.08	6.75	7		3	3	11	40	34	2	
C_1	5–2	3.85	39		1	1	4	5	48		1
	2–0.2	21.40	19		1		6		73		1
	0.2–0.08	13.95	7		3	2	5	13	67	4	
	< 0.08	1.65			5	5	15	25	50		

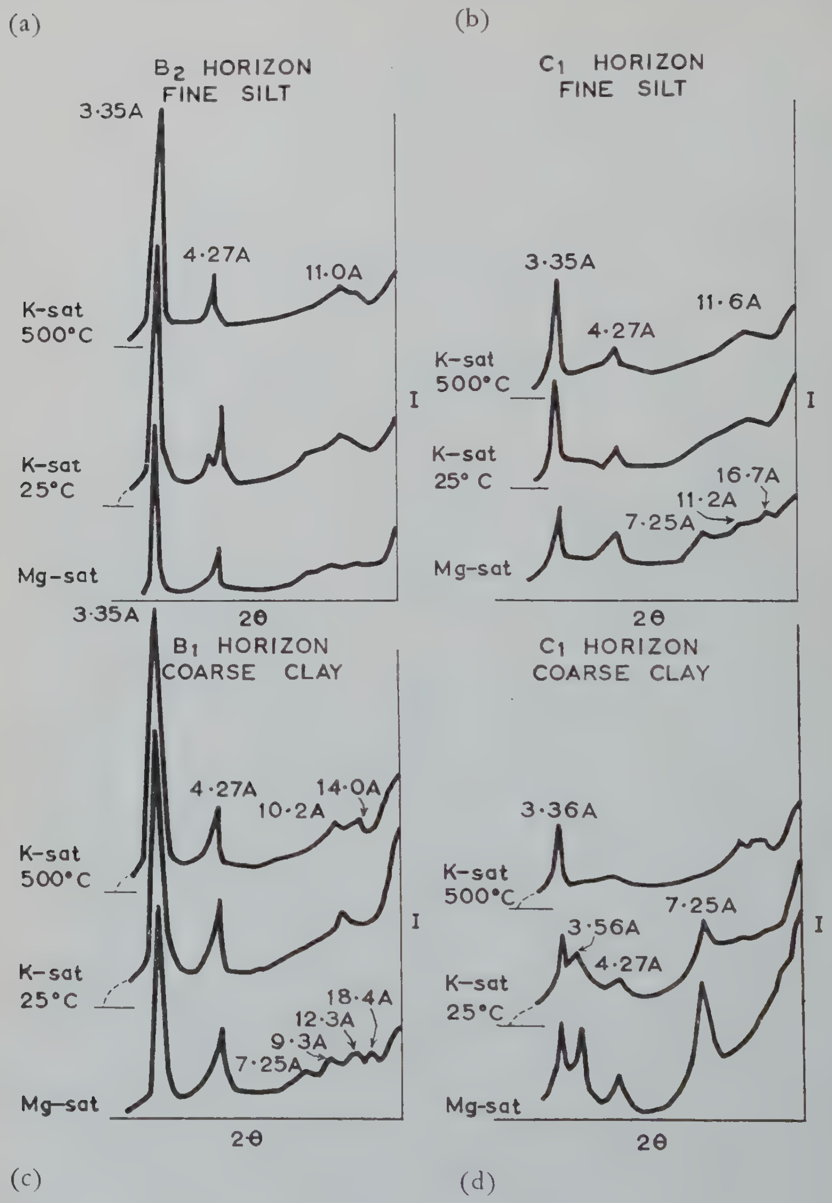

Fig. 4—X-ray diffraction patterns, obtained with basally oriented specimens, for the fine silt fractions of the Pukenamu B_2 and C_1 horizons, and the coarse clay fractions of the B_1 and C_1 horizons.

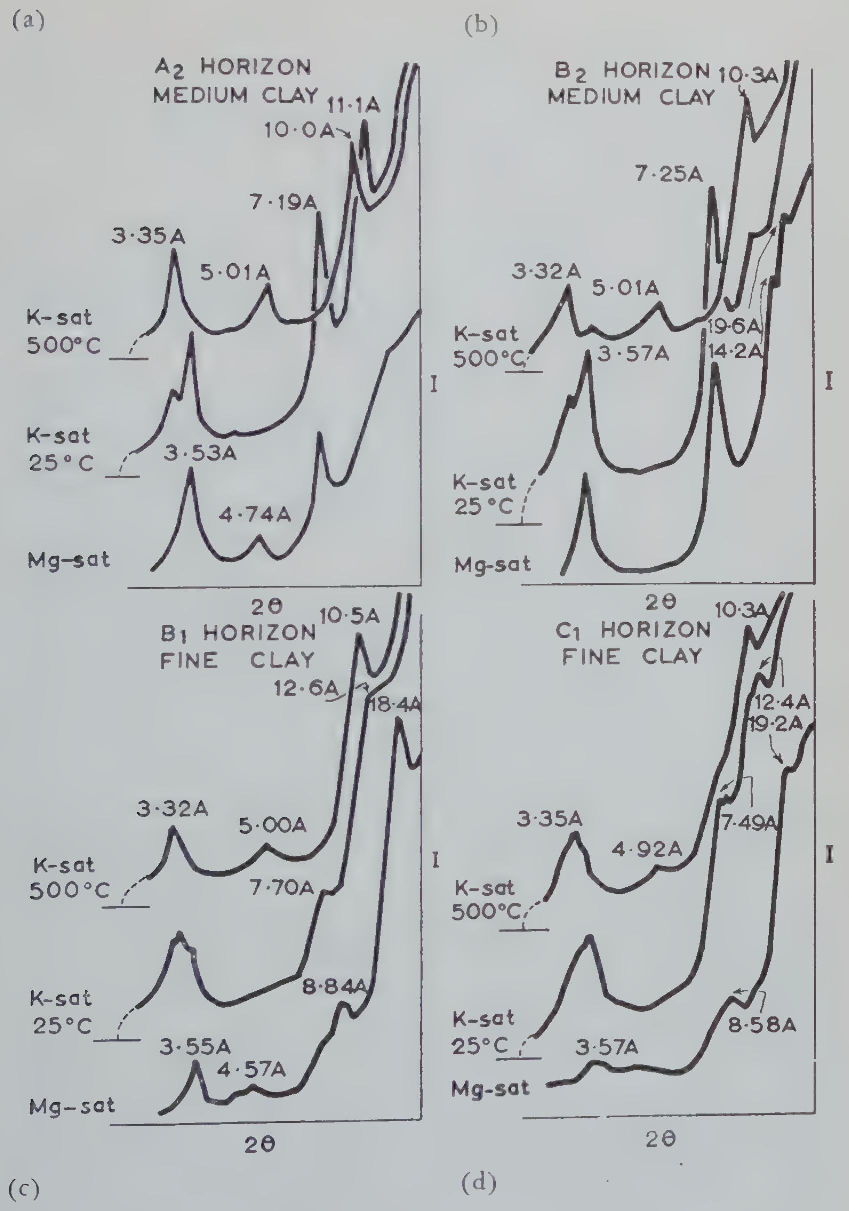

FIG. 5—X-ray diffraction patterns, obtained with basally oriented specimens, for the medium clay fractions of the Pukenamu A_2 and B_2 horizons, and the fine clay fractions of the B_1 and C_1 horizons.

Parahaki Sandy Loam

Parahaki sandy loam is similar to the Pukenamu silt loam in that it developed beneath a kauri forest. The Parahaki soil was strongly podzolised, however, because the forest was older and the rainfall in the region was higher than that of the Pukenamu soil.

The profile sampled had an A_2 horizon which was actually 15½ in. thick. The top 11 in. have been designated an A_1 horizon because of the assimilation of organic matter from existing scrub vegetation. Below the A_2 horizons there was a horizon from which much material had been eluviated, but in which humus had accumulated. It was therefore given the double symbol A_3–B_{21}. Beneath this were horizons of iron oxide accumulation (B_{22}), clay accumulation (B_{23}), and layers of partly weathered rock.

Mineralogy of the Sand and Silt Fractions: Quartz and cristobalite, sanidine, a few scattered grains of oligoclase, and small amounts of volcanic glass made up the sand fractions of this soil. In the C_1 horizon there was as much cristobalite as quartz but the cristobalite was a "stuffed" derivative structure which was more similar in its properties to a zeolite than it was to quartz. A discussion of this mineral will be found in the section on the fine silt fractions.

Quartz was the dominant mineral in the coarse and medium silt fractions down to the B_{23} horizon and the "stuffed" cristobalite referred to earlier was the dominant mineral in the C_1 horizon. The little feldspar found in these fractions seemed to have a chance distribution rising to 10% in the A_1 and B_{21} horizons. There may have been small quantities of interstratified layer silicates.

The fine silt fractions were similar to the other silt fractions in that quartz and cristobalite were the most abundant minerals in all horizons. Analytical data and mineralogical compositions for these fractions are shown in tables 7 and 8 respectively.

The fine silts of the A_1, A_2 and B_{21} horizons contained practically nothing but quartz and cristobalite, and chemical analyses showed 96–98% SiO_2 (see table 7). Cation exchange capacity, specific surfaces, and amounts of water present were all low. The X-ray diffraction patterns for the A_1 horizon, which have been chosen as typical, are represented in fig. 6a.

The fine silt of the C_1 horizon was also practically all quartz and cristobalite (see fig. 6d), but the analytical data for this fraction were very different from those of the three fractions referred to in the previous paragraph. This material contained only 62·85% SiO_2 and contained in addition 18·60% Al_2O_3 and a total of 10·95% water. The cation exchange capacity was over 100 m.e./100 g, and the specific surface was large. These differences are due to the type of cristobalite, which, in the terminology of Buerger (1954), is a "stuffed derivative" structure in which silicon is replaced by aluminium. Unlike the "stuffed" structures of Buerger however, this mineral did not have fixed cations to neutralise the charge unbalance caused by the substitution. Instead, the charge is neutralised by exchangeable cations and hydroxyl ions.

TABLE 7—Cation Exchange Capacities, Specific Surface Data, Data on Heating Weight Loss, and Chemical Analyses of the Fine Silt and Clay Fractions of Parahaki Sandy Loam

Horizon	Size Fraction μ	C.E.C. M.E. %	Specific Surface*	Heating-weight Loss 110–350° %	350–500° %	500–900° %	Chemical SiO_2 %	Al_2O_3 %	Fe_2O_3 %	MgO %	CaO %	Na_2O %	K_2O %	TiO_2 %
A_1	5–2	0.5	1.0	0.74	0.00	0.00	97.58	0.00	0.18	0.00	0.00	0.16	0.08	0.05
	2–0.2	6.0	2.2	0.40	0.40	0.00	96.05	0.89	0.38	0.07	0.00	0.20	0.05	1.13
A_2	5–2	1.6	1.0	0.15	0.00	0.10	96.90	0.00	0.19	0.00	0.00	0.24	0.12	0.58
	2–0.2	1.2	2.1	0.30	0.30	0.00	97.08	1.06	0.29	0.08	0.00	0.16	0.00	1.29
A_3-B_{21}	5–2	2.2	1.5	0.67	0.42	0.00	97.24	0.55	0.17	0.00	0.00	0.15	0.00	0.48
	2–0.2	2.6	1.5	0.67	0.42	0.00	89.46	6.00	0.51	0.31	0.18	0.35	0.11	2.33
	0.2–0.08		9.5	5.94	5.00	1.86	12.03	25.71	1.51	0.36	0.00	0.44	1.52	18.10
B_{22}	5–2	1.9	2.7	4.17	0.50	0.00	86.70	8.00	0.56	0.00	0.00	0.26	0.02	0.91
	2–0.2	5.0	5.9	8.38	2.02	1.81	59.68	27.40	0.94	0.24	0.00	0.10	0.14	1.68
	0.2–0.08	31.2	11.8				25.50	47.55	3.20	0.35	0.00	0.00	1.06	5.69
	< 0.08		25.6	10.7	6.1	6.1								
B_{23}	5–2	11.2	4.1	1.11	3.91	0.22	78.49	14.77	0.81	0.10	0.00	0.36	0.12	0.98
	2–0.2	23.3	6.8	2.73	8.61	1.96	51.33	33.65	1.27	0.25	0.00	0.10	0.20	1.15
	0.2–0.08			5.12	10.00	2.26	52.30	28.58	0.89	0.19	0.06	0.31	0.22	0.86
	< 0.08			7.4	9.5	3.7								
C_1	5–2	107	7.2	8.0	1.55	1.38	62.85	18.60	2.19	0.00	0.00	3.04	0.15	0.35
	2–0.2	106	29.0	10.81	2.76	2.50	52.84	27.77	1.65	0.15	0.00	1.13	0.36	0.43

*The units for specific surface are mg glycerol per 50 mg sample.

TABLE 8—Mineralogical Compositions of the Fine Silt and Clay Fractions of Parahaki Sandy Loam

Horizon	Size Fraction μ	Amount of Fraction in Horizon %	Qr %	Cb %	Mc %	Ch %	Vr %	Mt %	Kl %	Gb %	An %
A_1	5–2	8.10	60	37	1						1
	2–0.2	5.85	59	36		1			3		1
	0.2–0.08	0.15									
	< 0.08	0.00									
A_2	5–2	7.11	60	36	1						1
	2–0.2	5.79	60	36		1			2		1
	0.2–0.08	0.06									
	< 0.08	0.00									
A_3–B_{21}	5–2	6.00	60	37	1		1				1
	2–0.2	8.35	52	35	1	2	1		2	1	2
	0.2–0.08	1.30	12	10	13	4	1	5	22	16	18
	< 0.08	0.35	5		5	30	25	10	10	5	10
B_{22}	5–2	10.10	61	33		1			4	10	1
	2–0.2	11.85	41	10	2	1	1	2	15	26	1
	0.2–0.08	4.90	3		9	4	5		26	49	6
	< 0.08	1.00				15	15	5	27	35	
B_{23}	5–2	17.15	38	26	1	1	1		27	3	1
	2–0.2	16.55	14	9	2	2	2	4	54	7	1
	0.2–0.08	17.00	16		2		3	15	59	7	1
	< 0.08	0.95			10	10	10		60	10	
C_1	5–2	18.55	15	62	1	3	2		11		
	2–0.2	8.45	10	56	3	3	2		26		
	0.2–0.08	0.30									
	< 0.08	0.00									

In its properties the "stuffed" cristobalite was more like a zeolite or feldspathoid mineral than it was like quartz. It was possible, by calculation from the water loss and Al_2O_3 content, to obtain some measure of the amounts of this type of cristobalite in the horizons above the C horizon. In the fine silt fractions the mineral decreased very rapidly from 62% in the C_1 to less than 10% in the B_{23}, and there did not appear to be any in the upper horizons.

In the A_1, A_2, and A_3–B_{21} horizons there were only small quantities of interstratified layer silicates in addition to the silica minerals. The B_{22} horizon, however, contained 10% of gibbsite (see fig. 6b), and the B_{23} contained 27% of the same halloysitic kaolin mineral (see fig. 7a) which occurred in the lower horizons of the Maungarei and Pukenamu soils.

Mineralogy of the Clay Fractions: Analytical data for and mineralogical compositions of the clay fractions are shown in tables 7 and 8 respectively. Some representative X-ray diffraction patterns obtained with basally oriented samples are reproduced in figs. 6, 7, and 8.

The coarse clay fractions of the A_1, A_2, and A_3–B_{21} horizons were very similar to the corresponding fine silts in that they consisted almost entirely of quartz and cristobalite. X-ray diffraction patterns for the A_1 horizon are represented in fig. 6c. The patterns for the A_2 and B_{21} horizons were similar.

The coarse clays of the B_{22} horizon contained 15% of kaolin and 26% of gibbsite, although the silica minerals were still predominant. In the B_{23} horizon, however, the silica minerals made up only 23% of the coarse clay and there was 54% kaolin, 7% gibbsite, and lesser amounts of 2 : 1 and 2 : 2 layer silicates (see fig. 7b).

In the coarse clay of the C_1 horizon, quartz and cristobalite were again the predominant minerals although there was still 26% of kaolin (see fig. 8a). The cristobalite here was the "stuffed" variety containing large amounts of Al_2O_3 and water (see table 7). In the coarse clay fractions this type of cristobalite decreased sharply from the C_1 to the B_{23} horizon but small amounts of it appeared to persist up into the A_3–B_{21} horizon.

The medium and fine clay fractions of the two A horizons and of the C_1 horizon were practically non-existent, and in no horizon was there sufficient fine clay to allow chemical analyses to be made. The medium and fine clays examined showed similarities to the corresponding coarse clays but there were some differences.

The medium clay of the A_3–B_{21} horizon contained numerous secondary minerals. These were 13% mica, 22% kaolin, 16% gibbsite, and 18% anatase. The latter was identified in X-ray diffraction patterns (see fig. 8b) by a 3·50 Å maximum which persisted after the heating to 500°C had destroyed all the kaolin. The small amount of fine clay present also showed evidence of anatase (see fig. 8c) together with chlorite, vermiculite, montmorillonite, and kaolin. As the rhyolite contained only 0·6% TiO_2, and as no other fraction, with the partial exception of the medium clay of the B_{22} horizon, showed more than token amounts of TiO_2 in chemical analysis, it must be concluded that there had been a concentration of TiO_2 in these frac-

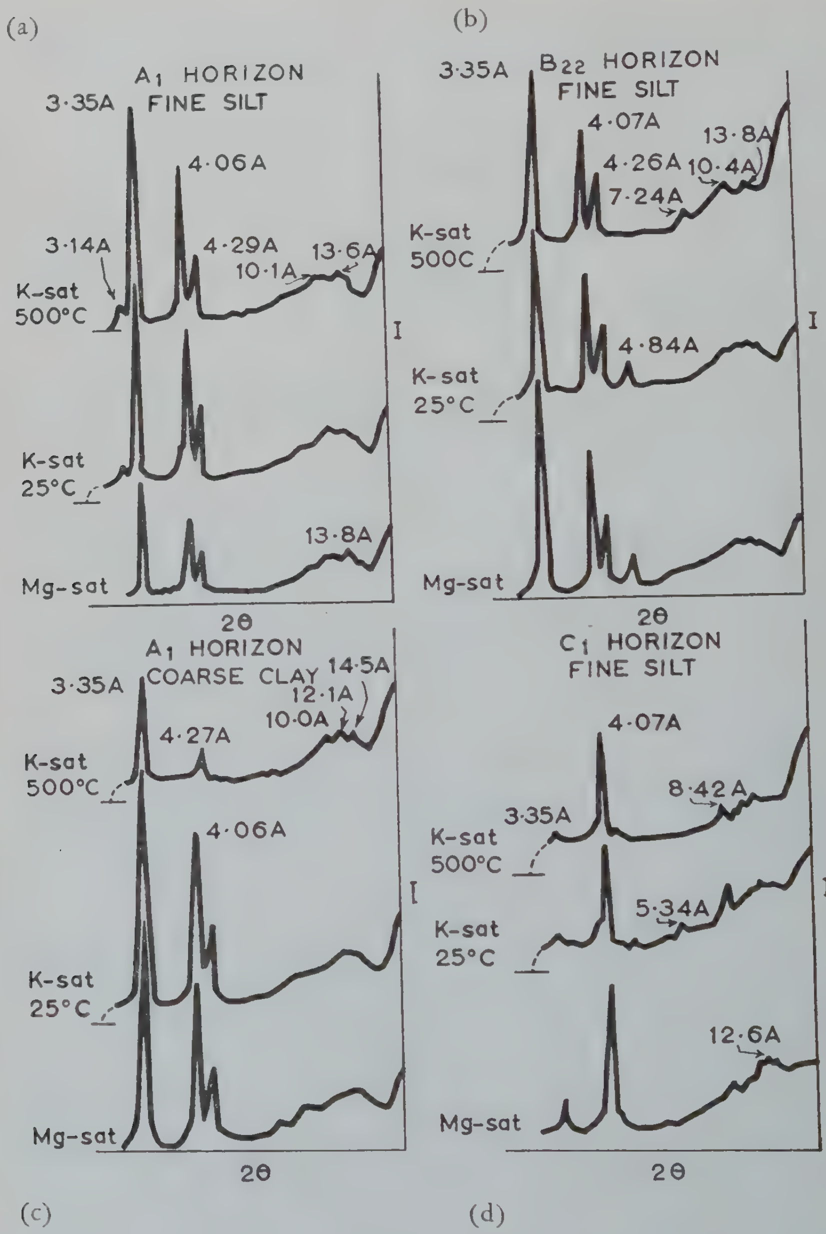

FIG. 6—X-ray diffraction patterns, obtained with basally oriented specimens, for the fine silt and coarse clay fractions of Parahaki A_1 horizon, and the fine silt fractions of the B_{22} and C_1 horizons.

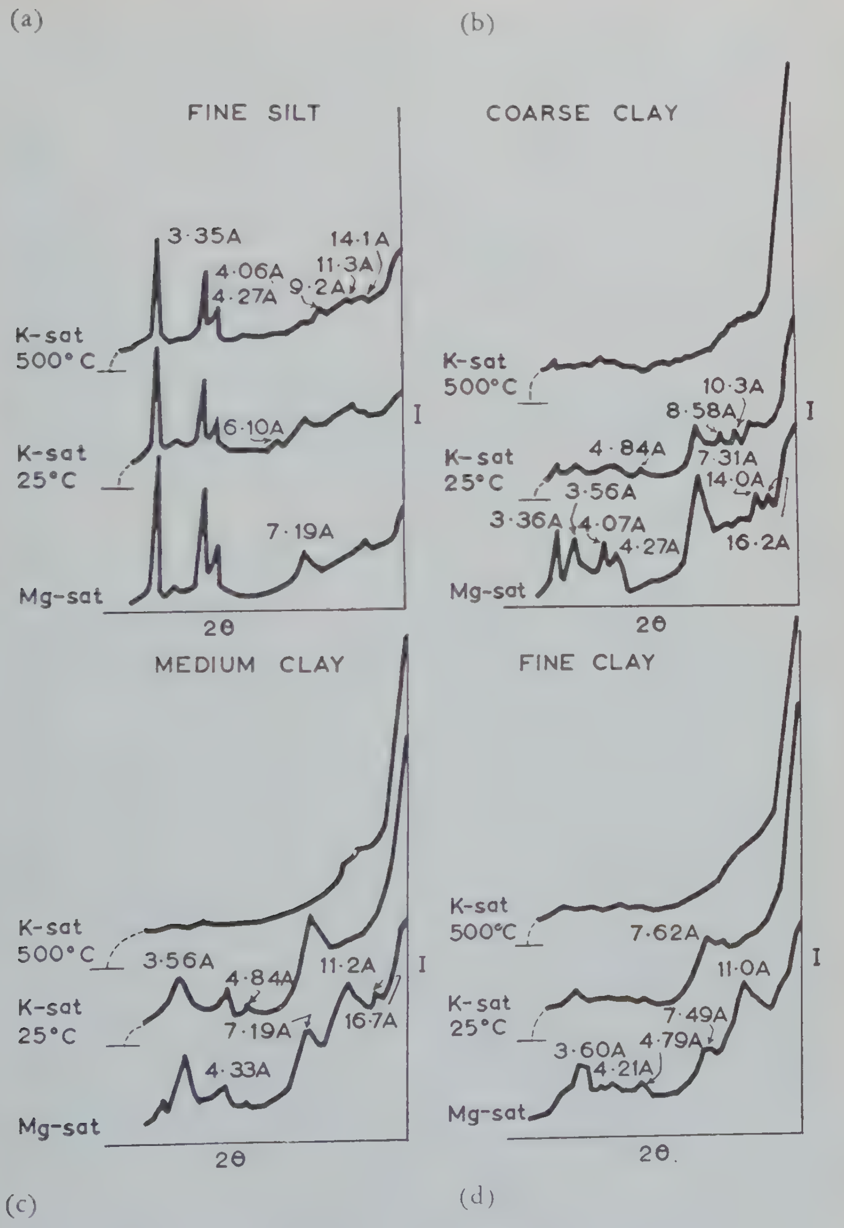

FIG. 7—X-ray diffraction patterns, obtained with basally oriented specimens, of the fine silt and clay fractions of the Parahaki B_{23} horizon.

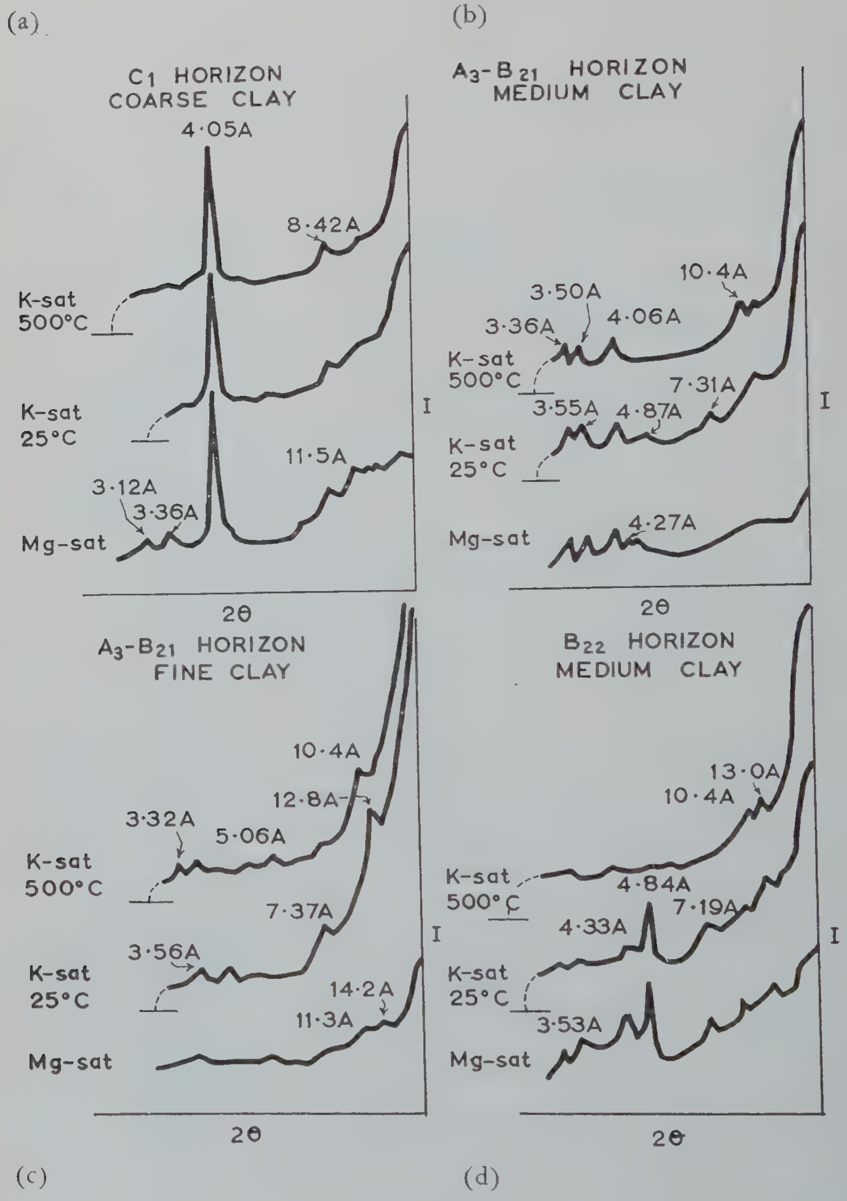

FIG. 8—X-ray diffraction patterns, obtained with basally oriented specimens, of the coarse clay fraction of the Parahaki C_1 horizon, the medium and fine clay fractions of the A_3–B_{21} horizon, and the medium clay of the B_{22} horizon.

tions. From its occurrence in the finer fractions, it seems likely that the anatase achieved its present position by a solution-precipitation mechanism affecting the over-lying horizons. Hence the A_3 B_{21} and to a small extent the B_{22} are illuvial horizons for TiO_2 in this soil.

The medium and fine clays of the B_{22} horizon were similar to their coarse clay in containing kaolin and gibbsite. As has already been mentioned, the medium clay also contained a larger amount of TiO_2 than the coarse clay. It is possible that the fine clay contained a relatively large amount of TiO_2 also, but there was insufficient fine clay to allow chemical analyses, and the X-ray diffraction patterns did not show anatase.

In the B_{23} horizon the medium and fine clay, like the coarse clay, was predominantly kaolin. The kaolin in these fractions appeared to be hydrated halloysite because the X-ray diffraction pattern (see fig. 7c) of the gycerol-solvated sample had a clear diffraction maximum at 11·2 Å whereas the K-saturated, non-solvated samples gave no maximum higher than 7 Å other than a small shoulder attributed to collapsed montmorillonite. A diffraction maximum at 4·33 Å with a profile characteristic of a two-dimensional reflexion was further evidence for the presence of halloysite.

DISCUSSION

The yellow-brown earths and podzolised yellow-brown earths are described by Taylor and Cox (1956). In these soils, weathering is considered to be a major soil-forming process, but because two of the soils are podzolised to a large extent, it is not possible to arrange the soils, from intrinsic evidence alone, in an order which shows the extent to which they have been weathered relative to one another, except that the Koikoi soil is clearly less weathered than the other three. Such an order can be obtained from considerations of the rainfall, temperature and degree of erosion at each site and of the length of time to which the rocks have been exposed to weathering. The order is Koikoi < Pukenamu < Maungarei < Parahaki.

Podzolisation has been impressed upon these soils by the organic cycle. From the general appearance of the profiles and from the known effects of the types of vegetation under which the soils have developed, the order of relative amounts of podzolisation is Koikoi < Maungarei < Pukenamu < Parahaki.

Thus the soils can be arranged in one sequence when the amounts of weathering they have undergone are considered, and another, different, sequence when the amounts of podzolisation are considered. It is of interest to examine the changes in particle size distributions, distributions of extractable iron oxides, and mineralogy in relation to these two.

Particle Size and Extractable Iron Oxide Distribution

The particle size distribution of the four soils show a trend towards greater dissimilarity throughout the profile as the soils become more podzolised (see table 2). The dissimilarities are the result of the decomposition and redistribution of clay by the processes of podzolisation. In the Pukenamu and Parahaki soils zones of clay accumulation occur in the subsoil.

In the Koikoi and Maungarei soils, where the amounts of clay decomposition and redistribution are small, the amount of clay is an indication of the amount of weathering, that is, there is more clay in the Maungarei than in the Koikoi soil.

There was little difference beween the extractable iron oxides in the two horizons of the Koikoi soil (see table 2). In the other three soils there were less extractable iron oxides in the topsoil than in the subsoil, the difference between the two increasing in the order of increasing podzolisation. In the Parahaki soil there was an horizon of iron oxide accumulation at the 19- to 24-in. depth.

Mineralogy of the Sand Fractions

The sand fractions by virtue of their grossness are more resistant to change than are the finer fractions of the soil. Because of this, variations in the mineralogical compositions of the sand fractions are dependent upon variations in the mineralogical compositions of the rhyolite from place to place. Comparisons of the mineralogy of the sands of the different soils are liable to tell more about the divergence of the rhyolites from the ideal of uniform composition than they tell about mineralogical changes due to katamorphism.

The mineralogical complements of the sands of all four soils were much the same, but there were faily large variations in the relative amounts of minerals, partly due to variations in the different rhyolites and partly due to katamorphic changes. There did not appear to be any trends related to katamorphism in passing from soil to soil. Within each soil profile, some trends related to katamorphism were seen. With depth, sanidine tended to increase relative to oligoclase, volcanic glass tended to increase in amount, and the amounts of silica minerals tended to decrease, particularly in the Parahaki soil. As these effects occurred in the weakly podzolised Koikoi soil they would seem to be primarily related to weathering. The pronounced increase in the amounts of silica minerals in the strongly podzolised Parahaki soil indicated that podzolisation had accelerated the trend towards silica mineral accumulation.

Mineralogy of the Coarse and Medium Silt Fractions

The mineralogical compositions of these two fractions are also dependent upon the mineralogical composition of the parent rock, although the dependence would be expected to be less than that of the sand fractions. This expectation was borne out by the results of petrographic, X-ray and cation exchange studies.

Only in the Koikoi soil were there appreciable amounts of volcanic glass and feldspars in the coarse and medium silts. These minerals occurred sporadically in the other soils. Quartz (plus cristobalite where it occurred) was always the dominant mineral, increasing in amount from the Koikoi to the Maungarei to the Pukenamu soil in which quartz made up more than 90% of the A horizon. The Parahaki soil contained over 90% of quartz and cristobalite in the A horizons.

Some aggregates of secondary layer silicates with 14, 10, and 7 Å spacings occurred in the coarse and medium silts of all soils. The amounts of layer silicates were greatest in the Koikoi soil and least in the Parahaki.

Within the profiles of the Maungarei, Pukenamu, and Parahaki soils the same trend of silica mineral accumulation occurred from the parent material to the surface of the soil. The amounts of other minerals, including the layer silicates, decreased in the same direction. The amounts of silica minerals were so high in the coarse and medium silts of these three soils that it was not possible to say to what extent the trend of accumulation was due to podzolisation and to what extent it was due to weathering.

Mineralogy of the Fine Silt Fractions

The upper size limit for fine silt is 5μ which is close to the upper limit of colloidal size. Fine silt particles are therefore more capable of reacting to a changing environment than either sand or coarse and medium silt fractions. For this reason, the mineralogical compositions of fine silt fractions depend more upon the bulk chemical composition of the parent rock than they do upon the bulk mineralogical composition.

Table 9 records the mineralogical compositions of the fine silt fractions of the four soils. In the Koikoi soil the fine silts contained about 40% feldspars. 30% quartz, and 30% secondary minerals of which more than 2/3rds were 2 : 1 and 2 : 2 layer silicates. In the Maungarei soil there were large amounts of quartz, and no feldspars, while kaolin comprised practically all of the secondary minerals. As the degree of podzolisation in these two soils was low, the great simplification in mineral complement and the increase in quartz content must both be attributed to weathering.

In the Pukenamu and Parahaki soils the upper horizons were almost devoid of minerals other than silica minerals, but in the lower horizons kaolin and, in one horizon, gibbsite appeared in appreciable amounts. As the Pukenamu soil is believed to be less weathered than the Maungarei soil, it is inferred that the greater amounts of quartz in the surface horizons of the Pukenamu were due to the increased degree of podzolisation. It has already been noted that the difference in degree of weathering between the Koikoi and Maungarei soils caused an increase in the quartz content, so that podzolic soil development has merely accelerated this trend.

Within the profile of the Koikoi soil the fine silts were very similar. The presence of 30% of secondary minerals indicated that some weathering had occurred in the soil. In the Maungarei soil the fine silts contained little else but quartz and kaolin, and as the quartz increased towards the surface the kaolin decreased. The difference between the quartz contents of the C_1 and A_3 horizons was so great, however, that it seemed difficult to believe that it was entirely the result of vertical weathering. The results tend, therefore, to indicate a greater amount of podzolisation in this soil than has been shown by any previous evidence.

In the Pukenamu and Parahaki soils, secondary minerals were practically confined to the lower horizons and silica minerals comprised nearly all of the fine silts of the upper horizons. The increase in silica minerals was obviously related to the amounts of podzolisation the soils had undergone. In the C_1 horizon of the Parahaki soil, cristobalite was the dominant mineral but it must be remembered that this was "stuffed" cristobalite which was more an alumino-silicate than a silica mineral.

TABLE 9—Mineralogical Compositions of the Fine Silt Fractions (5 to 2 μ)

Horizon	Sd %	Ol %	Qr %	Cb %	Mc %	Ch %	Vr %	Mt %	Kl %	Gb %	An %
Koikoi Fine Sand Loam											
A	21	22	31		5	5	4	4	8	2	
C	19	24	30		10	6	2		8		
Maungarei Clay Loam											
A_1			65	5	2	3	3		20	2	
A_3			65	5	2	3	3		25	2	
C_1			16	5	1	4	3		70		
Pukenamu Silt Loam											
A_2			90	3	2	2	2		3		1
B_1			94		1	2			3		1
B_2			84	2	1	1	1		10		1
C_1			39		1	1	4	5	48		1
Parahaki Sandy Loam											
A_1			60	37	1						1
A_2			60	36	1						1
A_3–B_{21}			60	37	1		1				1
B_{22}			61	33		1			4	10	1
B_{23}			38	26	1	1	1		27	3	1
C_1			15	62	1	3	2		11		

TABLE 10—Mineralogical Compositions of the Whole Clay Fractions ($< 2\mu$)

Horizon	Sd %	Qr %	Cb %	Mc %	Ch %	Vr %	Mt %	Kl %	Gb %	An %
				Koikoi Fine Sandy Loam						
A		20		28	8	11	7	21	5	1
C	3	14		24	8	13	9	24	2	1
				Maungarei Clay Loam						
A_1		14		1	3	4	1	75	3	
A_3		11		1	2	4	2	77	1	1
C_1		5			2	4	1	88		1
				Pukenamu Silt Loam						
A_2		70		1	1	5	8	14		1
B_1		63	1	4	1	6	11	13	1	1
B_2		42		3	1	6	13	32	2	
C_1		14		2	1	6	6	70	2	1
				Parahaki Sandy Loam						
A_1		59	36		1			3		1
A_2		60	36		1			2		1
A_3–B_{21}		45	34	3	3	2	1	5	3	4
B_{22}		28	7	4	3	3	2	19	33	2
B_{23}		15	4	2	1	3	9	56	7	2
C_1		10	56	3	3	2		26		

Mineralogy of the Clay Fractions

Clay fractions, because of their colloidal nature, have greater reactivities than coarser fractions. Because of this it is to be expected that clay fractions will reveal the most information about mineral changes due to katamorphism.

The clays of each soil were split, for purposes of analysis, into three size fractions. The mineralogical compositions of these fractions are given and commented upon in the results sections. Many of the layer silicates tend to concentrate in certain particle size ranges, and splitting the clays into three fractions often has the effect of separating complex mixtures into simpler, more easily determinable mixtures. For discussions relative to katamorphism, however, the mineralogical compositions of the whole clay fractions (less than 2μ equivalent diameter) must be considered. These compositions were obtained by multiplying the mineralogical compositions of the individual clay fractions by the amount of that fraction expressed as a percentage of the whole clay and summing the results for each horizon.

The results are shown in table 10. The virtual absence from the table of primary minerals other than silica minerals attests to the extent to which the compositions of the clays depend upon the katamorphic changes which have occurred.

The most obvious trend seen in passing from soil type to soil type down the table was the accumulation of silica minerals which took place in the eluvial horizons. The trend was obviously related to the degree of podzolisation, although weathering had undoubtedly contributed to it. The silica minerals almost certainly accumulated because the katamorphic processes were removing the other minerals either by decomposition and solution or by colloidal transfer. The fact that the trend was seen in the sand and silt fractions as well as in the clay suggests that decomposition and solution was the most important method of removal.

Some interesting variations in layer silicate contents can be seen in the table. In the Koikoi soil the clays contained appreciable amounts of mica, chlorite, expanding layer silicates, and kaolin. Kaolin minerals made up almost all of the layer silicates of the clays of the Maungarei soil. The clays of the Pukenamu soil also contained large amounts of kaolin but they contained significant amounts of montmorillonite and vermiculite as well. In the clays of the Parahaki soil the amounts of 2 : 1 and 2 : 2 layer silicates were small (with the partial exception of the B_{23} horizon), and kaolin was the predominant layer silicate. Gibbsite appeared in appreciable amounts in the B_{22} horizon.

In the weathering sequence theory presented by Jackson and Sherman (1953), it was suggested that a succession from micaceous minerals (mica and chlorite) to expanding layer silicates (vermiculite and montmorillonite) to kaolin to gibbsite appeared in soils subjected to increasing degrees of weathering. A succession of this nature can be seen in the present set of soils when they are placed in the order Koikoi, Pukenamu, Maungarei, Parahaki, which is the order of increasing weathering suggested by considerations of rainfall, temperature, and degree of erosion. The authors are of the opinion, therefore, that these results give confirmation for soils developed from a single parent material to stages 7 (mica) through 11 (gibbsite) of the weathering sequence theory. Stated otherwise, the qualitative mineralogical compositions of the clays of any three of these soils from rhyolite could have been predicted with a high degree of success from a knowledge of three things, viz. (a) the weathering sequence theory, (b) the mineralogical compositions of the clays of the fourth soil, and (c) the relative degrees of weathering of the four soils. This statement does not imply any continuity between the stages of the weathering sequence, that is, it is not implied that a clay which now consists almost entirely of kaolin (say) has necessarily consisted, at some time in the past in the same environment, of montmorillonite or of mica.

Within each soil profile the trend of quartz accumulation was well in evidence, particularly in the two more podzolised soils, suggesting again that the trend was more directly related to podzolisation than to weathering. In the Pukenamu soil the quartz appeared to accumulate more at the expense of kaolin which was reduced fivefold between the C_1 and A_2 horizons than at the expense of montmorillonite and vermiculite which remained at approximately the same percentages. If it were possible to generalise from such scanty evidence it would seem that kaolin is more susceptible to decomposition by podzolisation than is montmorillonite or vermiculite. Although the evidence is scanty, the generalisation is logical enough because of the known ability for podzolisation to mobilise aluminium. If the destruction of the minerals is due to the mobilisation of their aluminium, kaolin, having numerous aluminium ions on the surface of the layer lattice, would be more easily decomposed than montmorillonite and other 2 : 1 layer silicates where much of the aluminium is held in octahedral positions inside the lattice. Carrying this thought further we would expect gibbsite to be even more susceptible to attack than kaolin so that in this respect the processes of podzolisation tend to be the reverse of the processes of weathering. Rode (1944), after extensive review of the literature on podzolisation came to a similar conclusion. He states: "the most characteristic feature of the processes (of podzolisation) is the circumstance that even the secondary clay minerals are subject to decomposition. These minerals are the products of weathering that takes place in the thermodynamic systems of the surface of the earth and as such should be more stable. Their instability in the process of podzolisation illustrates the principal difference between the processes of weathering and soil formation."

The quartz which accumulated was, in all probability, primary quartz derived from the parent rock. The cristobalite which accumulated along with quartz in the eluvial horizons of the Parahaki soil may well have been a secondary mineral produced by soil forming processes. The cristobalite in the C horizon of this soil was "stuffed" cristobalite containing approximately 25% Al_2O_3 and 10% water. The cristobalite in the eluvial horizons was practically pure silica so that even if it formed directly from the other cristobalite, a large amount of ionic rearrangement must have occurred.

The clays of Koikoi, Maungarei and Pukenamu soils did not show any variations with depth which could be used as evidence for the progression of weathering in a vertical direction. In the Parahaki soil the predominance of kaolin in the B_{23}, gibbsite in the B_{22}, and the relatively high amounts of anatase in the A_3–B_{21} might be evidence for a tendency for minerals more resistant to weathering to become more prevalent as the regions of highest weathering intensity at the soil surface were approached. The extent to which this trend is confounded by the precipitation of ions and minerals swept out of the eluvial horizons is unknown. It may be large so that the results must be interpreted with caution.

FORMATION OF THE SOILS

From the evidence collected in this study, an explanation of the manner in which the soils were formed can be offered. To make this explanation it must be assumed that the effects of ecosystems which existed prior to the ones observed have been obliterated. This assumption, questionable for most soils, is probably valid for the Koikoi soil because of its azonality and degree of erosion, and for the other three soils, because of the strong weathering of the present regime and, where it occurs, of the effect of the kauri.

A further assumption made for clarity and convenience in discussion is that weathering has proceeded prior to and separately from podzolisation. In as much as weathering, like laterization, is in its strictest sense a process of soil-material formation rather than soil formation (see Byers, Kellogg, Anderson, and Thorp, 1938, p. 974), this assumption may actually represent the truth for a period of time following the exposure of the rock at the earth's surface. For most of the time, however, the two processes will act contemporaneously and their effects will be concurrent rather than consecutive.

Koikoi Fine Sandy Loam

The Koikoi soil was formed under a low to moderate degree of weathering. The rhyolite was changed to a clastic mixture of primary and secondary minerals, the secondary minerals being synthetised from the initial decomposition products of the primary minerals according to the amounts of silica, alumina, and K and Mg available. The availability of these substances was determined by the amounts in the parent rock, the rate and order of primary mineral decomposition, and the rate of leaching. In the Koikoi parent material, sufficient K and Mg were retained to allow mica, chlorite, and the expanding 2 : 1 layer silicates to form, but there was not sufficient to prevent the formation of kaolin and gibbsite as well. Most of the 2 : 1 and 2 : 2 layer silicates were discrete and presumably were formed separately, but X-ray diffraction revealed some interstratification of mica, vermiculite, and chlorite. This may suggest that one of these, e.g., mica, was formed first and the others were formed from it, *in situ*, by further weathering. This implies that the vermiculite and chlorite were dioctahedral. But our knowledge of the conditions of formation of these minerals is incomplete, and the possibility that the interstratified mixtures crystallised directly from the initial decomposition products must not be excluded.

For some reason iron oxides are taken up by layer silicates only in minor amounts. Most of the iron oxides which were liberated in the decomposition of the primary minerals were disseminated throughout the material as hydrated oxides which gave the material its light brown colour, and contributed to the formation of blocky structure.

The secondary minerals contained, in total, less silica, K, Ca, Na, and Mg than the parent rock. Silica, K, etc. not used in the formation of these minerals were leached out, and there was a nett loss of material from the site. Primary quartz, which was virtually unaffected by weathering, increased in amount proportionally to the other minerals.

Although weathering was more intense at the surface, the position of the site on a steep, eroding slope counteracted the effects of vertical weathering.

Under the beech forest which clothed the soil about 100 years ago there may have been a low degree of podzolisation which caused some slight redistribution throughout the profile. With the removal of the beech and the subsequent burning and overgrazing of the tussock which replaced it, the soil quickly eroded. The A horizons were stripped off, the horizons were mixed by soil creep and frost action, and a light brown, shallow, stony, undifferentiated soil, which contained some primary and some secondary minerals and more quartz than the rhyolite resulted.

Maungarei Clay Loam

The Maungarei clay loam was formed under a moderate-to-high degree of weathering, in which the primary minerals other than quartz were almost all decomposed. The rate of leaching was high, and kaolin was formed in quantity from the decomposition products of the primary minerals. Kaolin was almost the sole secondary mineral because it was nearly in equilibrium with the combination of weathering environment and rhyolite rock. This equilibrium was indicated in the results by the lack of both 2 : 1 layer silicates and gibbsite. Consequently the material was composed almost entirely of kaolin and quartz, and continued weathering tended mainly to increase their amounts.

Iron oxides liberated by primary mineral decomposition were hydrated and disseminated throughout the material, giving it a uniform, yellowish brown colour. The nett loss of material from the site was high, and the proportion of quartz left was much greater than in the rhyolite. Because of the preponderance of quartz, the texture of the material was primarily dependent upon the natural particle-size distribution of the quartz, and secondarily dependent upon the particle-size distribution of the kaolin. As the rock was microcrystalline, the quartz was distributed from fine sand to coarse clay. The kaolin was mainly fine silt and clay, and the combination gave a silty clay loam material.

The permeability was not good, and water was held up in the lower levels. The environment in these levels was slightly wetter than at the surface, and the kaolin minerals formed were more hydrated than at the surface.

A low degree of podzolisation mobilised the iron oxides in the surface levels, and caused them to be carried to a depth at least greater than 30 in. The material from which the iron was removed became paler in colour, but the colour difference was obscured by the assimilation of strongly-pigmented organic matter in the upper levels. Some of the kaolin was also decomposed and removed, and the proportion of quartz was increased. As the kaolin was removed, the percentage of clay decreased and the textures of the eluvial zone become coarser than the texture of the rest of the material. The result was the Maungarei soil.

Pukenamu Silt Loam

The Pukenamu soil was formed under a moderate-to-high degree of weathering, under which the non-quartz primary minerals were almost all converted into secondary minerals with a large nett loss of material. Leaching removed most of the K, Mg, Ca, and Na from the weathered material, and the predominant secondary mineral was kaolin. The removal of K was more nearly complete than the removal of Mg, however, and the amount of mica was much less than the amounts of montmorillonite and vermiculite. These minerals occurred in interstratified minerals in which they may have formed.

The total amount of iron in the rhyolite was low (ca 2%), and some of the iron liberated during decomposition of the primary minerals was taken up in the expanding layer silicates. The remainder appeared as hydrated iron oxides which gave the material a pale yellowish brown colour.

The texture was dependent primarily upon the particle-size distribution of the preponderant quartz, and secondarily upon the particle-size distribution of the kaolin and other secondary minerals. The quartz, owing to the fine-grained nature of the rhyolite, was mainly silt, and the secondary minerals were fine silt and clay size, with the vermiculite and montmorillonite concentrating in the fine and medium clay. The textures were therefore silty clay loams. Prior to the onset of podzolisation there was probably more kaolin and correspondingly less vermiculite and montmorillonite at the surface than there was at depth. In the lower levels a restriction in the percolation rate provided a wetter environment than at the surface and the kaolin was more halloysitic than at the surface.

This material was subjected to a moderate degree of podzolisation under kauri vegetation. Iron oxides were mobilised and removed almost completely from the surface and less completely from the lower levels. They were carried down at least beyond 40 in. They left a pale grey to pale brown material which was further changed by the decomposition and removal of much of the kaolin previously present, and of some of the montmorillonite and vermiculite as well. The removal of this secondary material made the texture a silt loam. Some of the eluviated minerals were reformed and precipitated at depths below 10 in., and the clay contents in these levels rose accordingly. A soil with a light-grey, quartzose A_2 horizon, a transitional B_1 horizon, and a clayey, yellow-brown B_2 horizon containing quartz, kaolin, and expanding layer silicates was the result.

After the kauri was removed by man, scrub vegetation grew up in its place, and the soil was modified to some extent. The scrub vegetation of New Zealand produces very darkly-pigmented humus, and the assimilation of this in the soil obscured the grey colour of the podzolised A_2 and darkened the B horizons as well, without having any noticeable effect on the texture or mineralogy of the soil.

Parahaki Sandy Loam

The Parahaki soil was formed under a high degree of weathering, under which primary non-quartz minerals were converted principally into gibbsite and kaolin. Physical breakdown of the rock produced coarse sand particles which resisted weathering more successfully than their finer counterparts, so that moderately resistant minerals such as sanidine were able to persist in small quantities even in the most weathered regions near the surface. The rate of leaching was very high, and so much silica was removed, that gibbsite was formed in quantity, although there was sufficient silica to permit the formation of moderate amounts of kaolin. The nett loss of material from the site was also high, and primary quartz, which was resistant to weathering, was the predominant mineral left. Iron liberated from the primary minerals was precipitated as hydrous oxides which were concentrated as other materials were removed by leaching. Titanium released from the weathering minerals was precipitated as fine-grained anatase and was also concentrated by the nett loss of material.

The texture of the material was dependent primarily upon the particle-size distribution of the quartz, and secondarily upon the particle-size distribution of the secondary minerals. The quartz ranged from coarse sand to coarse clay, and the secondary minerals were mainly fine silt or finer. The resulting texture was a silt loam.

Owing to the attenuation of weathering with depth, the amount of gibbsite decreased and the amount of kaolin increased with depth. The nett loss of material was lessened – and partly alleviated – by the deposition of material from above, and the concentrations of quartz, iron oxide and anatase were reduced at depth. The influence of quartz upon the texture was diminished and the texture became finer with depth, moving into loam and clay loam classes. Eventually a level was reached where there was a fall off in the amounts of secondary minerals and the textures were coarsened once again.

A high degree of podzolisation was imposed upon this material. Under its influence the upper regions were virtually depleted of all their secondary minerals, and some of their persistent non-quartz primary minerals were also removed. An eluvial zone was produced in which quartz, as the only mineral resistant to both weathering and podzolisation, was heavily concentrated. The small amounts of primary minerals which, because of their coarseness, had resisted weathering were also concentrated in the eluvial zones. Secondary silica crystallised as cristobalite was produced in moderate quantities in the eluvial zone. The eluvial zone therefore consisted of loose, incoherent grains of quartz, cristobalite, and some other primary minerals. The amount of clay was small, and the texture was sandy. The zone, at the time the kauri was removed, extended to a depth of 18½ in.

Because the podzolising agents lost their effectiveness with depth some of the ions and minerals they transported out of the eluvial zone were precipitated within the material, and a zone of illuviation was formed. The permeability of the silt loam was probably low, and the zone of illuviation was formed, initially, fairly close to the surface. Continued action of the podzolisation process gradually caused the eluvial zone to deepen, and the illuvial zone receded from the surface. At the time the kauri was removed, the illuvial zone occupied a position from 15 to around 40 in. and consisted of three horizons: B_{21} from 15 to 19 in. in which TiO_2 and humus matter were deposited; B_{22} from 19 to 24 in. in which gibbsite and iron oxides were deposited and formed – the latter particularly in a 10-mm. indurated layer at 19 in.; and B_{23} from 24 to around 40 in., in which kaolin was deposited and formed. The horizon from 15 in. to 19 in. was an eluvial horizon in its principal mineralogy, texture, and structure, and an illuvial horizon in its accumulation of humus matter and TiO_2.

Large amounts of fine-grained minerals were deposited or formed in the B_{22} and B_{23} horizons. Increased clay contents were thereby superimposed upon a textural gradient which passed through a fineness maximum somewhere above the 40 in. depth. The resulting textures were loams and clay loams, and the maximum clay content occurred in the B_{23} horizon.

When the kauri was removed, scrub vegetation took its place. Humus, assimilated by the podzol from this vegetation, coloured the top 11 in. dark grey. Some of this humus was carried down by water into the A_3–B_{21} horizon where it was deposited on top of the indurated iron layer. It darkened this horizon, making the accumulation of humus in this horizon much more obvious than it had been under the kauri. Strictly, therefore, the podzol is a relict profile and the soil-forming processes under the present vegetation acting on the sandy loam A horizons of the podzol are causing the soil to tend towards the intrazonal group.

Mechanisms of the Genetic Processes

In explaining the geneses of these soils, the authors have avoided as much as possible any statement of the mechanisms by which the processes acted. The number of contributing mechanisms may have been very large indeed. They may have been physical, such as insolation, abrasion or comminution; or chemical, such as solution, chelation or reduction. Some of these mechanisms can be related to the process of soil parent material formation and others to the process of soil formation, but there are some which belong in either category. In fact, the separateness between soil parent material and soil becomes exceedingly hard to maintain at the mechanistic level, and it is more accurate to consider all these mechanisms, collectively, as mechanisms of katamorphism.

Swindale and Jackson (1956) explained some of the mineralogical results obtained in this study in terms of two katamorphic mechanisms which they called soluviation and cheluviation. Soluviation and cheluviation were actually compound mechanisms because they combined the physical mechanism of eluviation with the two chemical mechanisms of solution and chelation. They applied only to the decomposition aspects of katamorphism, and the synthetic aspects were achieved by subsequent, unstated means. The two mechanisms were theoretical, evolved from considerations of the ways in which complex silicate minerals decompose in water. Their application to these soils was made mainly to indicate how these theoretical mechanisms were thought to act in practice.

The explanation seemed to fit the facts so well that it is possible to assert that they probably were the principal mechanisms of decomposition in these soils. If this is true, it implies that the main illuvial mechanisms were precipitative, with cations and anions deposited at levels determined by the changing chemical environment. Synthesis of new minerals then occurred by crystallisation from the precipitated materials and from existing residues of prior decompositions.

CONCLUSIONS

A mineralogical examination of four yellow-brown earth and podzolised yellow-brown earth soils derived from rhyolite has revealed much evidence concerning mineralogical changes attendant upon weathering and upon podzolic soil development, and has allowed a logical explanation of the formation of the soils to be advanced.

In the soils the minerals varied in their powers of resistance to weathering in the following order: oligoclase < sanidine < mica (muscovitic) = chlorite < expanding micas (vermiculite and montmorillonite) < kaolin < gibbsite. Rhyolitic volcanic glass was less stable than sanidine, and quartz was at least as stable as gibbsite. The secondary minerals included in this order were formed during weathering as less resistant minerals decomposed, and in the clay-size fractions of the soils a part of the weathering sequence put forward by Jackson and Sherman (1953) was recognised.

There was some evidence to suggest that the order of resistance of minerals to podzolisation was almost the reverse of the order of resistance to weathering, but again quartz was very resistant. Secondary silica, crystallised as cristobalite, appeared to be the only mineral formed by podzolisation, the main role of which seemed to be otherwise confined to the decomposition of the minerals formed by weathering.

When weathering and podzolisation both acted upon the soil material, the most obvious effect was the rapid accumulation of residual quartz in the surface horizons.

The sand fractions of the soils contained only primary minerals. They indicated the relative stabilities to weathering of these primary minerals.

The coarse and medium silt fractions were composed almost entirely of quartz, with only small amounts of other primary minerals and secondary minerals. Because of the preponderance of quartz, they were of little use in indicating katamorphic trends in the soils.

The fine silt fractions contained secondary minerals in large quantities. They also contained much quartz, and in the less weathered soils, other primary minerals. They illustrated better than the other fractions the transitions from primary to secondary minerals, and the relative stabilities of the most stable primary minerals and the least stable secondary minerals.

The clay fractions contained quartz and secondary minerals. They best illustrated the relative stabilities of these minerals, and the trends due to increasing degrees of weathering and increasing degrees of podzolisation. There were evident tendencies for montmorillonite to concentrate in the fine clay ($< 0{\cdot}08\mu$) and for kaolin and gibbsite to concentrate in the medium clay ($0{\cdot}2$–$0{\cdot}08\mu$).

The geneses of the four soils may be summarised as follows:

Koikoi sandy loam was formed by a low to moderate degree of weathering with little or no podzolisation. The soil was composed of quartz, feldspars, volcanic glass, and some secondary minerals, mainly mica and kaolin. There was no profile differentiation shown by particle-size distribution analysis, by extractable iron oxide contents, or by mineralogy.

Maungarei clay loam was formed under a moderately high degree of weathering with a small degree of podzolisation. The soil was composed of quartz and kaolin with small amounts of other primary minerals and even smaller amounts of other secondary minerals. There were gradual increases in amounts of clay, of extractable iron oxides, and of non-quartz minerals with depth.

Pukenamu silt loam was formed under a moderate to high degree of weathering with a moderate degree of podzolisation. The soil showed marked differentiation of the profile. The eluvial horizon consisted almost entirely of quartz. The illuvial horizons contained quartz, kaolin and significant amounts of montmorillonite and vermiculite. They contained more clay and more extractable iron oxides than the eluvial horizon.

Parahaki sandy loam was formed under a high degree of weathering with a high degree of podzolisation. The soil showed very marked profile differentiation. The eluvial horizons were composed of quartz and secondary cristobalite, and contained little clay and practically no iron oxides. The illuvial horizons contained minerals highly resistant to weathering (quartz, kaolin, gibbsite, anatase), and were high in clay and extractable iron oxides. Three zones of accumulation were recognised in the profile. These were: B_{21}, humus and anatase; B_{22}, iron oxides and gibbsite; B_{23}, kaolin. The C horizon contained an unusual "stuffed" cristobalite with high cation exchange capacity, high specific surface, and a large water-holding capacity.

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CLAY MINERALS, ACCELERATED EROSION, AND SEDIMENTATION IN THE WAIPAOA RIVER CATCHMENT

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Summary

In the Waipaoa River catchment, clay minerals were used to identify the source rocks contributing appreciable quantities of fine sediment to the floodplains of the Lower Waipaoa River valley. The conclusion arrived at is that all easily erodible rocks contribute material and that sediment is not all derived from rocks showing spectacular forms of erosion. The clay mineral, montmorillonite, is present in all rocks of the catchment and is partly responsible for the high degree of erosion in the crushed argillites at the headwaters of Waipaoa River.

INTRODUCTION

In parts of the Waipaoa River catchment there is serious accelerated erosion and during major floods, appreciable quantities of fine sediment have been deposited on floodplains of the lower Waipaoa River valley. One problem was to name the source rocks from which the sediment was derived. Because flood deposits are mainly fine sands, silts, and clays it had proved impossible to identify sources merely on lithology; a possible alternative was to use clay minerals by comparing the assemblage in easily erodible rocks of the uplands with that in soils of the floodplains. Another problem was to examine the extent to which clay minerals are responsible for the highly erodible properties of crushed argillites.

Waipaoa River Catchment

The Waipaoa River is one of the two principal rivers in the Gisborne - East Coast region. Its tributaries arise on the eastern slopes of the Raukumara range and the river flows in an almost southerly direction to enter the sea near Gisborne (fig. 1). The area of the catchment is more than 500,000 ac. The land is, for the most part, moderately steep, but there is flat to easy rolling land on high level terraces flanking the main valley from Gisborne to beyond Whatatutu. The lower Waipaoa River valley or Gisborne Plains has an area of 50,000 ac and the land is the most highly productive in the region.

Rocks

Rocks are principally of Tertiary age with some Cretaceous. Formations have been described by Henderson and Ongley (1920), and surface rocks by Grange and Gibbs (1948), and Hamilton and Kelman (1952). In their report, Hamilton and Kelman classified the rocks in order of increasing erodability, but in this paper, only easily erodible rocks are considered, viz. crushed mudstones, bentonitic mudstones, and crushed argillites. The forms of erosion occurring on them are briefly described below:

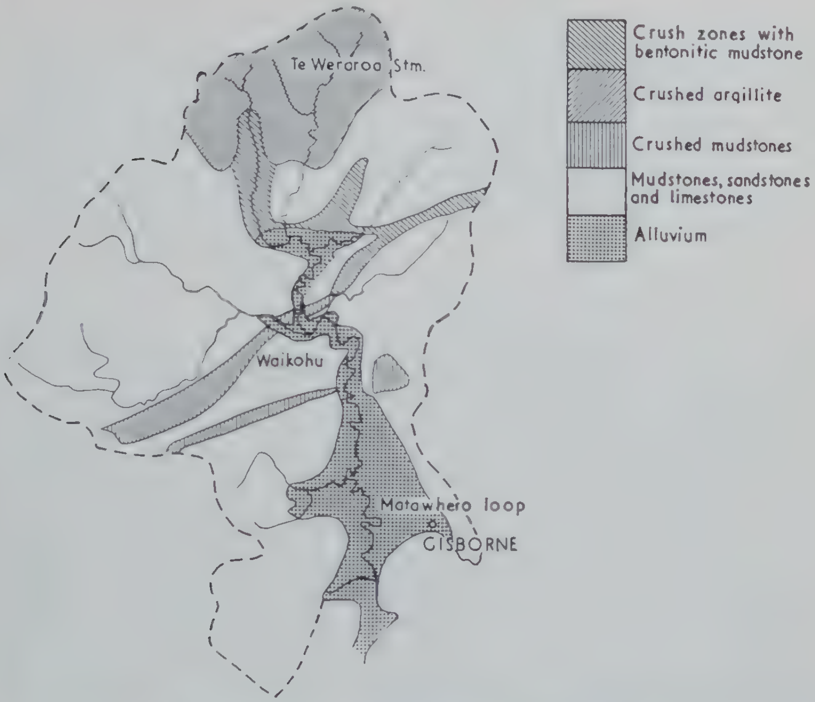

FIG. 1—Simplified geological map of the Waipaoa Catchment (after Hamilton and Kelman).

(i) In crushed mudstone (Tertiary), erosion is predominantly sheet with shallow slips.

(ii) Bentonitic mudstones (early Tertiary), which lie in long, narrow crush zones, have a hummocky and churned-up surface resembling that of a glacier. When saturated with water, large areas slump and flow even at angles of only a few degrees. This form of erosion is not as spectacular as that in crushed argillites because slumped areas do not usually lose their plant cover. These rocks contribute much colloidal material to flood sediments.

(iii) In crushed argillites (Cretaceous), erosion takes the form of large gullies which erode rapidly headwards. From these gullies, enormous quantities of fine, easily transportable rubble and mud is carried down the side of the gully by rainwater and then poured into river channels, causing them to aggrade quickly.

ACCELERATED EROSION

Accelerated erosion is severe in the northern part of the catchment. Attention has been drawn to it from time to time by Henderson and Ongley (1920), Cotton (1948), Grange and Gibbs (1948), and Hamilton and Kelman (1952). Many unpublished reports have also been made on aspects of erosion and erosion control by the New Zealand Soil Conservation and Rivers Control Council and the Poverty Bay Catchment Board. River-bed measurements indicate that in the middle reaches, the bed has risen 3 ft in the last eighty years and 20 ft and more in the upper reaches. Measurements of sediment deposited on the Gisborne Plains since 1932 show that in low-lying parts near Waipaoa River, the build-up is quite appreciable. To protect farming land on the Plains, an expensive flood control scheme involving stop-banks and cut-off channels in the river, is now well under way.

SAMPLES

The samples studied are shown in table 1. Some amplification of the description of the sites may be necessary. Erosion is most spectacular in the areas of crushed argillite and bentonitic mudstones, and especially is this so in the Te Weraroa stream (fig. 1). In the large gullies formed by rapid erosion, blue grey argillite (sample 6934) is seen to be overlain by a layer about twenty feet thick of brown argillite, (6936) which although crushed seems to be relatively resistant to erosion. Within the blue argillite are hard bands of brown rock (6935) which also resist erosion and stand out on an eroding face of the blue argillite. These appear to be lithological, whereas the brown argillite follows the topography and appears to be the result of weathering of the blue argillite, causing oxidation of the iron. This brown or weathered argillite erodes only slowly, but when gullies in it cut through to the blue argillite below, then erosion proceeds very rapidly, undercutting the brown argillite above.

On the plains, samples were taken from deposits laid down by a recent flood (August 1956, samples collected in September 1956, sample 6932), and from an older alluvium exposed by excavation for river diversion work at Matawhero bridge. This alluvium is a clay, overlain by debris derived from the Taupo ash showers of about 1800 years ago, and was sampled to gather information on the character of river deposits at that time.

It was also possible to make use of results of clay mineral analyses of recent soils of the Gisborne Plains analysed by L. D. Swindale.

EXPERIMENTAL METHOD

Samples 6928 – 6931, having dried and hardened, were ground to pass a 200-mesh sieve, other rock and sediment samples were broken up while still wet, while soil samples were air dried and ground to pass a 2-mm sieve. Free iron oxides were removed from some samples and estimated using Aguilera and Jackson's method (1954). Organic matter was removed by oxidation with hydrogen peroxide, and the clay and silt fractions separated, using sodium hydroxide at pH 10.5 as the dispersing liquid. Although the percentages of sand, silt and clay so obtained will not necessarily be equivalent to those obtained by standard methods of mechanical analysis, they give an indication of the texture of the material and form a useful basis for the comparison of samples.

TABLE 1—Samples Studied

Lab. No.	Depth	Description	Locality	Soil Type
6270A	0–4 in.	Black sandy clay loam	Haisman Rd., Hexton	Kaiti sandy clay loam
B	9–15 in.	White sandy-clay matrix	,,	,,
C	15–18 in.	Pale yellow sticky clay	,,	,,
6272A	0–3 in.	Black clay	Muriwai	Kaiti clay
B	13–18 in.	Light olive grey clay		,,
6271A	0–4 in.	Very dark grey clay loam	Makauri	Makauri clay loam
B	12–16 in.	Grey mottled clay		
6928	12–24 in.	Weathered crushed argillite	Te Weraroa Exptl. area.	
29	24 ft	Brown hard crushed argillite	Te Weraroa Exptl. area.	
30	50 ft	Shattered blue argillite	Te Weraroa Exptl. area.	
31	50 ft	Slickensided calcareous mudstone	Waerenga-o-Kuri.	
32	0–6 in.	Dark silty clay	Matawhero loop cut-off	(Flood deposit, August 1956)
33	10 ft	Brown mottled clay	Matawhero loop cut-off	(Under stump and pumice bed)
34	40 ft	Blue shattered argillite	No. 24 Gully, Te Weraroa Exptl. area.	
35	30 ft	Brown hard mudstone band in blue argillite	No. 24 Gully, Te Weraroa Exptl. area.	
36	5 ft	Weathered crust below pumice cover	No. 24 Gully, Te Weraroa Exptl. area.	
37	3 ft	Bentonitic mudstone	Tarndale Rd., Mangatu.	
38	3 ft	Brown clay	Waikohu	Slightly weathered tertiary mudstone

In order to aid quantitative analysis of the clay fractions, some of these were separated into four size fractions; 2–1μ, 1–0·2μ, 0·2–0·1μ, and less than 0·1μ. It was found that most of the quartz and feldspar in the clay fractions was concentrated in the 2–1 fraction, while the 0·1 fraction was nearly all montmorillonite, with a little illite. It was possible to make an approximate analysis of samples not fractionated by comparison with samples whose analysis was known by fractionation.

Then clay fractions were studied by conventional X-ray spectrometer and differential thermal analysis techniques. Calcium carbonate was estimated by the method given by Metson (1956). A spectrographic analysis of some of the samples was carried out using a Hilger large quartz spectrograph.

RESULTS

The results of these analyses are shown in tables 2 and 3. The sedimentary rocks examined vary widely in their particle size distribution, but not in their mineralogical content. The finest textured rocks are the bentonitic mudstones (samples 6931 and 6937), the latter having 70% less than two microns and 35% less than 0·1 micron. The shattered argillites contain 50% or more of sandsize material, while the Tertiary mudstone is intermediate in grade.

TABLE 2—Mineralogical and Mechanical Analysis

	Sand (%)	Silt (%)	Clay (%)	Fe[1] (%)	Fe[2] (%)	$CaCO_3$ (%)	Qtz[3] (%)	Feld-spar[3] (%)	Illite[3] (%)	Mont^{m}[3] (%)	Kl[3] (%)
6270A	53	23	24								
B	41	26·3	33								
C	16	37·5	47								
6271A	10	42·0	48·5								
B	23	38·7	38·1								
6272A	20	42·7	37·2								
B	4	42·4	54·1								
6928	33·5	34·5	32				15	10	25	50	......
29	50·5	23	26·5				15	10	20	30	15
30	45	23	32				20	15	20	40	5
31	4	7	89				6	6	10	50	16
32	35	27	38	0·73	5	3·9	5	3	17	44	......
33	1	29·5	69·5		5		12	12	19	48	10
34	60	22	18	0·28	2		15	15	20	45	5
35	88	8	4	0·58	2						
36	75	16	9	0·49	2						
37	7	20	73	0·18	5	8·6	......	......	12	86	2
38	25	31·5	43·5	1·18	7		5	5	36	57	......

[1] Iron extracted by Aguilera and Jackson's method (1953).
[2] Total iron by spectrographic analysis.
[3] As percentage of total clay fraction.

TABLE 3—Mineralogical Composition of the Fractionated Clays

	Amount of Fraction (%)	Size Range (μ)	Illite	Montmor-illonite	Kaolin	Quartz	Feldspar
6931	34	2–1	20	20	20	20	20
	20	1–0·2	33	33	33	......	......
	4	0·2–0·1	20	60	20	......	......
	42	<0·1	10	80	10	......	......
6932	17	2–1	30	......	......	25	25
	18	1–0·2	40	60	......	......	......
	10	0·2–0·1	20	80	......	......	......
	25	<0·1	......	100	......	......	......
6933	10	2–1	25	25	......	25	25
	29	1–0·2	30	50	20	......	......
	8	0·2–0·1	20	70	10	......	......
	14	<0·1	5	90	5	......	......
6937	12	2–1	50	50	......	......	......
	28	1–0·2	25	70	5	......	......
	11	0·2–0·1	......	100	......	......	......
	49	<0·1	......	100	......	......	......
6938	16	2–1	40	......	......	30	30
	30	1–0·2	60	40	......	......	......
	7	0·2–0·1	40	60	......	......	......
	47	<0·1	20	80	......	......	......

The clay mineralogy is similar for all these rocks, the dominant clay mineral being montmorillonite, with illite and vermiculite also present. Quartz and feldspars are present in the coarser part of the clay fraction, and make up most of the sand and silt fractions. Shards of volcanic glass were observed in the sample of Tertiary mudstone (6938). Only the bentonitic mudstone contained any free calcium carbonate.

The two samples from Matawhero bridge (6932–33) also differed markedly in particle size distribution, the older one being practically all clay (70%), while the recent sample contained 35% sand. Mineralogically they were similar, containing large amounts of montmorillonite with illite, vermiculite, quartz and feldspars in lower amounts. Free calcium carbonate was observed only in the most recent sample (6932)—about half the amount found in bentonitic mudstones.

Extractable iron oxides, a figure which includes iron weathered out of iron-containing minerals and deposited as a coating of hydrated ferric oxides on the clay particles, show some significant variations. The total iron content of the argillite samples 6934–6936, as shown by spectrographic analysis, is the same, yet the extractable iron content of the blue, highly erodible argillite is about half that of the brown weathered argillite, and that of the resistant brown bands in the blue argillite. Probably the iron oxides have some effect in cementing the material together, thus making it more resistant to erosion.

Clay mineral analysis of recent soils (Swindale, pers. comm.) show that these contain mostly montmorillonite, illite, and vermiculite with iron oxides and small amounts of quartz and feldspar.

DISCUSSION

By using the above analyses, it was not possible to trace the source of any particular material brought down to the plains. Montmorillonite is found not only in the bentonitic mudstones, but also in the crushed argillites and in the Tertiary rocks. It seems likely that the Tertiary rocks have been derived in part from the erosion of Cretaceous rocks containing a high proportion of montmorillonite.

The present flood deposits are derived from both the Cretaceous and Tertiary rocks, but as these have similar mineralogy, minor differences must be used to estimate the relative importance of these rocks as sources of debris. The coarse texture of the recent flood deposit, sample 6932, shows that a large proportion of this material is derived from the rapidly eroding argillite, but the high calcium carbonate content shows that considerable quantities of the bentonitic mudstones are involved, although some of the calcium carbonate could have been derived from limestones in the Tertiary sequence. Erosion is not very severe on these limestones however.

The coarse nature of sample 6932 compared with the very much older 6933 shows that there has been a very considerable increase in the proportion of sand brought down by the river in the last two thousand years. However, the soils of the plains form a sequence of various ages up to about two thousands years old and they are all very much finer than the recent alluvium except for one horizon of Kaiti silty clay which is more sandy. This is thought to be the oldest soil on the plain and represents a period of rapid erosion at the beginning of the present cycle when the bush cover of the hills was destroyed in places by the Taupo ash shower.

Clay mineral examination amply confirms the surface observations of Grange and Gibbs (1948), and Hamilton and Kelman (1952) that soil erosion from rocks like crushed mudstones, while not spectacular, is steady and continuous. Such erosion is of concern to soil conservationists who must have regard for sustained agriculture i.e. the farmer must adopt measures to keep the soil in place.

The reason for the accelerated erosion is not hard to find. The argillite is very much shattered and broken, and has little coherence. It contains quite a high proportion of montmorillonite which is easily released by light grinding such as would take place when the rock was shattered and now fills the gaps between the shattered blocks of argillite. This acts as a cement when the rock is dry but as a lubricant as soon as it becomes wet. When water penetrates the rock, the montmorillonite swells and forces the blocks apart, rendering a moderate slope very unstable. When even a little material is removed from the base of the slope by stream undercutting the whole mass slumps by deep-seated rotational slips and flows at very small angles. All such movement grinds more montmorillonite out of the rock fragments and further lubricates the mass of debris.

The reason for the stability of the superficial oxidised zone is the cementing action of the iron oxides which coat the clay minerals, prevent the expansion of montmorillonite and lock the whole mass together. Under natural conditions the angle of slope of the land was much steeper, because of the presence of this oxidized zone. Only when this oxidized zone is broken by shallow gullying and undercut by floods caused by increased runoff, is the blue mudstone exposed, and because of its much lower angle of repose it begins to erode rapidly.

CONCLUSION

There are no marked mineralogical differences between the different rock types in the catchment area that can be used to determine the origin of the debris in the river. The predominance of montmorillonite is unexpected but accounts for many of the features of the erosion pattern of the area. No evidence has been found for the origin of the montmorillonite in crushed argillites but the montmorillonite in the Tertiary and recent beds is probably derived from erosion of Cretaceous rocks.

The high total iron content of the plains sediments, (samples 6932 and 6933), shows that a considerable portion of these materials has been derived from the Tertiary mudstones, containing iron which is probably of volcanic origin. The calcium carbonate content shows that a considerable contribution is being made by the calcareous bentonitic mudstones, as the Tertiary crushed limestones are not subject to much erosion, while the coarse sandy material is derived from the shattered argillite.

There is an increasing amount of coarse material in the recent deposits, which means that the crushed argillites are supplying a much higher proportion of sediment to the plains than herefore. These are not the only sources of debris, however, and erosion must also be checked on the mudstone regions where erosion is not so spectacular but is nevertheless important.

ACKNOWLEDGMENTS

The samples were collected with the aid of Mr W. A. Pullar, Soil Bureau pedologist, Gisborne, and Mr H. S. Gibbs, senior pedologist, Wellington. Thanks are due to Mr Pullar for making available background information and to Dr L. D. Swindale of the Soil Bureau for making available clay mineral analyses of soils of the Gisborne Plains.

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NOTES ON THE STRUCTURE OF GREYWACKES AND ARGILLITES AT TAWHARANUI PENINSULA, AUCKLAND

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(Received for publication, 11 September 1959)

Summary

Statistical analysis of bedding surfaces, shear surfaces and joint planes indicates that a basement series of greywackes and argillites has undergone at least two phases of folding of contrasting styles. It is suggested that an earlier phase of large-scale flexural slip-folding about a N.N.W. axis, β, along with attendant overturning, was followed by a phase of folding parallel to B induced by movement on oblique shear surfaces; the B axes are marked by corrugations. An anticlockwise rotation of approximately 30°, bringing β to the present N.W. trend during overturning, is suggested to explain its relation to the N.N.W. trend of B.

INTRODUCTION

The following account results from part of research undertaken for a thesis in support of the degree of M.Sc. at Auckland University College in 1955.

The area of investigation lies approximately 70 miles north of Auckland by road (fig. 1) and the basement rocks, exposed there under a cover of Tertiary beds of the Waitemata Group, are typical of the other basement rocks found on the coast and offshore islands along the eastern side of the Auckland Peninsula.

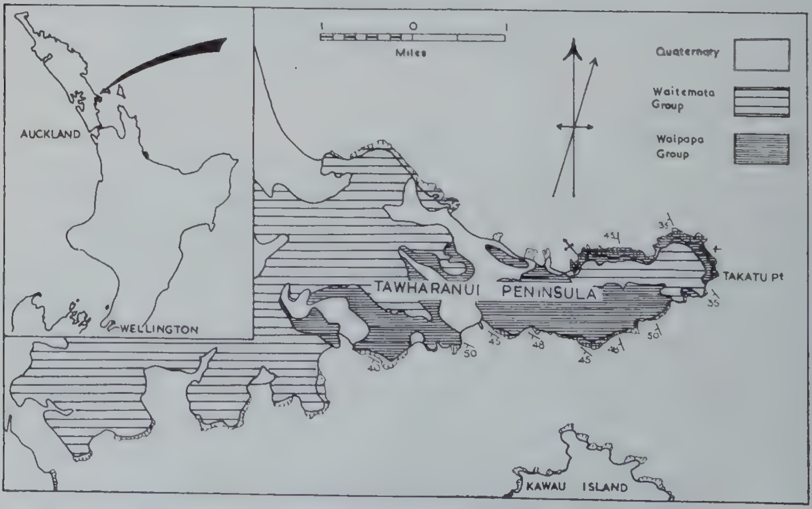

FIG. 1—Map of Tawharanui Peninsula showing outcrop of the Waipapa Group. Inset shows location of field.

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The strata, belonging to the Waipapa Group of Mesozoic age (Ferrar, 1934), consist of indurated greywacke sandstones and argillites and differ notably from sandstones of comparable age on the west coast of the North Island in that they bear the imprint of a long and complex tectonic history. The evidence of considerable deformation appears in large shear surfaces, abundant joints and general contortion of strata; the grade of metamorphism is equivalent to subzone Chlorite 1 of the Otago schists (Brothers, 1956).

Characteristically the strata show great apparent irregularity of dip and strike; shearing-out and lensing of the sandstones which alternate with argillites is a common feature. A complex joint pattern further complicates the picture so that earlier workers in the field experienced considerable difficulty in portraying the structure of the Waipapa Group except in a very generalised manner (Ferrar, 1925, 1934; Bell and Clarke, 1909).

The present discussion, based on field work on Tawharanui Peninsula and using the methods of Sander (1934, 1950), is an attempt to shed further light on the tectonic history of the area. It stands as supplementary to similar and more comprehensive work by Brothers (1956).

STRUCTURE

Within the basement group, two broad lithological types are recognisable: massive sandstones in thick layers; and alternating thin beds of sandstones and argillite. The first of these – the massive sandstones – comprises grey and

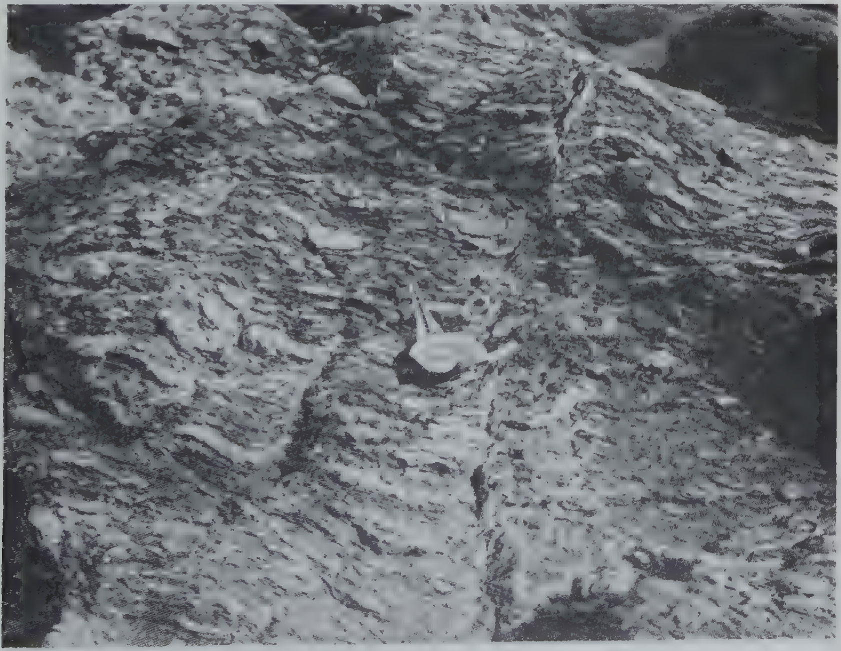

FIG. 2—Sheared alternating argillites and sandstones intersected by S_2, showing corrugations (B) and drag on ends of sandstone boudins.

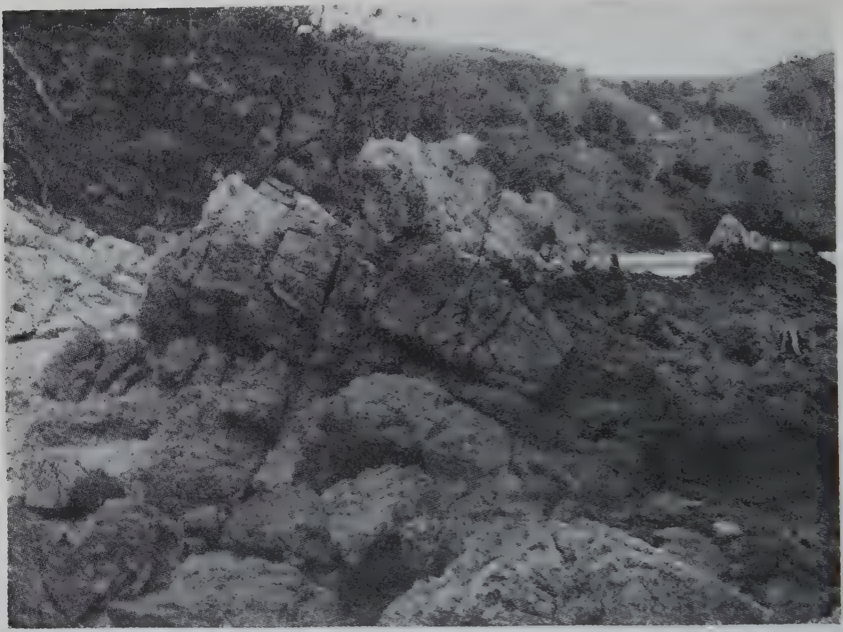

FIG. 3—Jointing S_3 in massive greywacke sandstones.

pale-green highly indurated greywacke sandstones, of very diverse grain size ranging from fine sandstones to sharpstone conglomerates containing fragments of argillite a centimetre or more in length. These massive beds are much fractured by regular joint planes often sealed by quartz veins, and by widely spaced zones of sheared and mineralised rock, but they do not exhibit any other macroscopic evidence of deformation.

The alternating sandstones and argillites occur in bands normally varying from ½ in. to 4 in. in thickness and seldom exceeding 6 in. in width. In contrast to the massive sandstones, the alternating sequences show several features indicative of mechanical deformation including small-scale folds, shear surfaces and boudinage. Only rarely do thick argillite successions constitute important lithological groups.

Structural data have been obtained from the alternating sandstones and argillites since these have reacted more readily to deforming stresses than have the massive sandstones. The thinner bands of sandstones in the alternating sequence have readily responded to deformation by folding, as well as by shearing and jointing, since the intervening layers of argillite have acted as a "lubricant", facilitating bedding-plane shear during flexural slip-folding. Concomittant oblique shearing has frequently resulted in extreme boudinage of the competent sandstones so that the original beds are now represented by discontinuous lenses and sheets isolated in the less competent argillite. Where oblique shearing and boudinage are most marked, small corrugations a few inches in wavelength are present and these lie at an angle to the general dip of the major stratification. The wavelength and amplitude of these folds varies considerably and folds as much as a foot across have been seen in the field, whilst others are so small as to be discerned only in thin section.

Before plotting of the data, areas of exposed greywacke were arbitrarily subdivided into approximately equal sectors and measurements taken from each of these were recorded separately.

The following structural elements were measured and plotted statistically as lower hemisphere projections on a Schmidt equal area net and the plot was then contoured on the basis of percentage per one per cent area.

1. Bedding, S_1.
2. Oblique shears, S_2.
3. Joints, S_3.
4. Small folds and corrugations, B.

The symmetry of the resultant plots was then considered in relation to the geometry of the structures observed.

1. S_1, originally a bedding surface, has become a surface of shear in the argillites in response to deformative stress. Initial anisotropy provided by the alternation of relatively competent and incompetent layers was such that the sequence was readily adaptable to adjustment by flexural slip-folding. The competent sandstones were thus usually able to fold concentrically between the intervening argillites which absorbed the bedding plane shear, but in some places shearing was of such intensity that the finer sandstone laminae in the argillites were completely shattered. Shearing oblique to S_1 has also taken place in the argillites and has strongly affected the enclosed sandstones.

Evidence of stratification in the massive sandstones is at best displayed as only a faint banding due to slightly differing grain size and fine pebble and intraformational conglomerates. Within these massive rocks bedding-plane shear is absent and structural adjustment is not by folding but is entirely restricted to movement along widely-spaced, large-scale oblique shear surfaces.

2. Throughout the Group, large zones of shear transect rocks of all types, including the massive sandstones, and these zones appear as bands, several inches to a foot wide, of sheared and crushed epidotised rock veined with quartz. Juxtaposition of different lithologies and pronounced slickensiding indicate that considerable movement has taken place along these fractures, but the extent of this movement is not determinable because of the lack of strata distinctive enough to serve as marker beds.

Evidences of shearing parallel to these zones, but on a much smaller scale, are seen in hand specimen as small surfaces of differential movement, and as rucking and boudinage of sandstone bands; even smaller-scale equivalents of these are recognisable in thin section as minute shear surfaces standing oblique to sub-microscopic laminae in the argillites. S_2 on all scales stands almost invariably at high angles (30°) to the bedding.

3. Surfaces of fracture distinguished from S_2 by lack of displacement are collectively designated as joints (S_3) and are recorded separately. The pattern and style of jointing in the basement rocks is apparently very complex and variable in expression. In the sandstone members S_3 is as a rule spaced widely and regularly and may be either open or sealed by quartz veins. In the argil-

lites, on the other hand, although normally not well developed, S_3 is sometimes expressed as a series of close-spaced parallel hair-line fractures. In both lithologies, variability in attitude of S_3 has resulted in fracturing of the rocks into rhomboidal blocks.

Although longitudinal (h0l) joints are present, most of S_3 surfaces measured fall into the category of cross or (0kl) joints and these are steeply inclined to the horizontal.

The likelihood of finding curved joints formed by flexure of a set of earlier S_3 was borne in mind, but no such structures were seen in the field. Hence, either the observed joints represent the final phases of deformation in the basement rocks, or later deformation was not accompanied by appreciable flexure.

4. Intimately associated with the small-scale shearing are corrugations (B-axes) of average wavelength from six inches to a foot, and amplitude approximately one third of this; however, in some instances, folds parallel to B have been seen with wavelengths exceeding three feet. The axial surfaces of folds are normally close to vertical but in a few cases are overturned and even recumbent. Generally, however, the ultimate strength of the sandstone has been exceeded before this stage of folding has been reached and the characteristic arrangement of curved, lens-shaped boudins in a sheared argillite matrix results. Overturned and recumbent folds generally occur only among the B-folds of shorter wavelength and seem to have formed through a combination of drag and small-scale overthrusting on the shear surfaces.

The corrugations and boudinage so formed have been exposed by marine erosion as a strong B-lineation plunging oblique to the dip of the macroscopic S_1 surfaces

DISCUSSION

Throughout, the field S_1 preserves a generally south-western dip and is not obviously thrown into large-scale folds. However, overturning, determined from depositional features preserved in the thicker argillites is apparent at several places where as many oriented specimens as possible were collected in order to determine the direction of younging of the strata. Graded bedding, penecontemporaneous micro-faulting and load casting were used and general agreement was found between the evidence of the different features for each of the individual localities.

It was found that the strata are all inverted except for a small section at the west end of the north coast. A noteworthy feature of the strata at Takatu Point on the eastern tip of the Peninsula is that they are considerably less sheared and corrugated than those elsewhere. Also, the small-scale folds are very much shallower and less distinct.

From consideration of the plots from individual sectors it was immediately apparent that the lack of inland exposure in the field was going to present a serious drawback to detailed structural work of a statistical nature. Although data were collected from a considerable length of shoreline, the width of exposure is limited as a rule to only a few yards. For this reason plots obtained from different sectors did not give strong maxima and the results from the sub-areas were therefore combined. It is apparent from the composite diagrams that a certain degree of inhomogeneity* exists but it is considered to be only slight and, bearing this factor in mind, some significant points emerge which are worthy of consideration.

In spite of the appearance of a general monoclinal dip of S_1, a β-plot of this surface revealed a maximum (β) plunging at 10° to 320°. The attendant partial girdle is considered to arise from spread resulting from the combining of several sectors. (fig. 4).

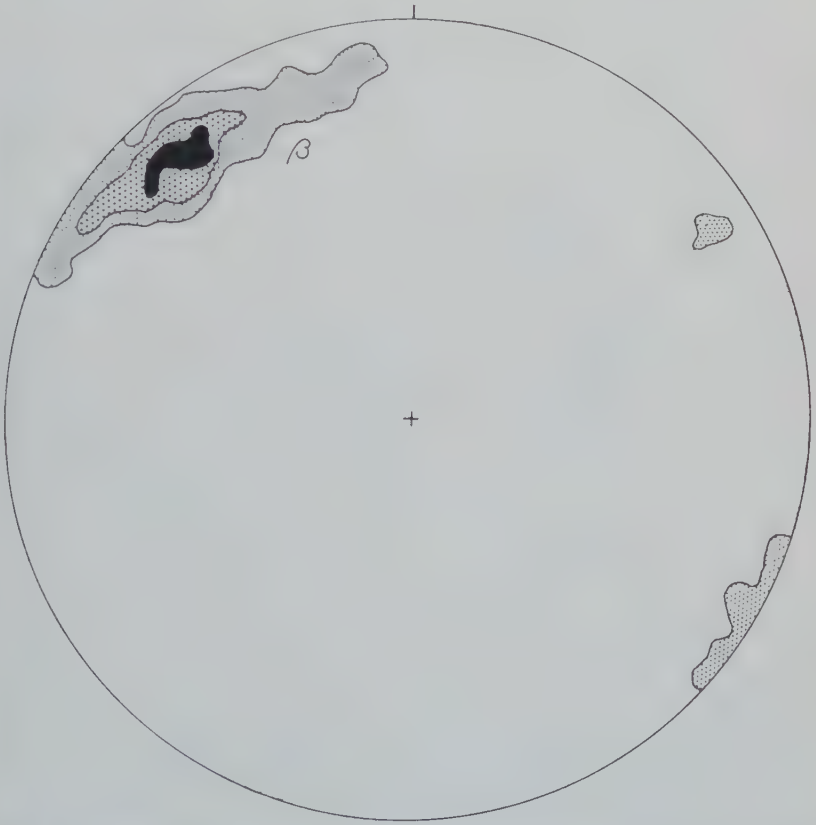

FIG. 4—β-diagram of S_1 based on 164 points. Contours at 6, 9, 12% per 1% area.

Contouring of β-points obtained from the intersection of plotted S_2 disclosed two distinct maxima plunging at 10° to 315° (β'_1) and at 50° to 150° (β'_2). Again partial girdles are present (fig. 5).

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*Weiss (1957, p. 587) makes the following statement concerning homogeneity; "A given field is most commonly considered structurally homogeneous if the form surface is either planar in orientation or folded cylindroidally about a single, statistically defined fold axis (β)".

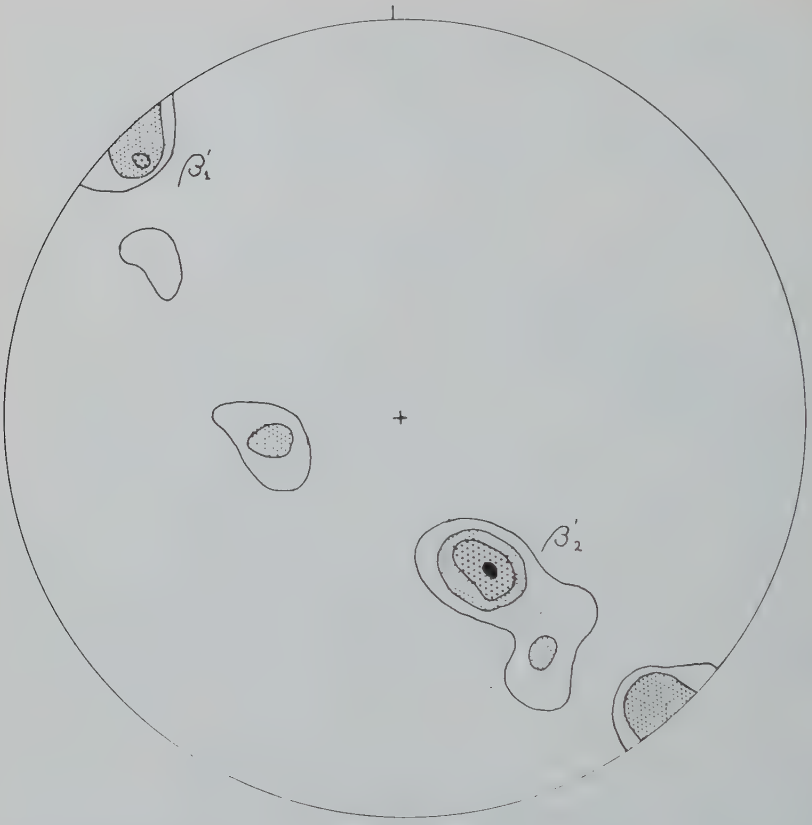

FIG. 5—β-diagram of S_2 based on 253 points. Contours at 4, 6, 8, 10% per 1% area.

Plotting and contouring of poles to S_3 (π–S_3) showed one maximum plunging at 5° to 335° and another plunging at 35° to 130° (fig. 6).

B-axes plotted and contoured showed a maximum plunging at 45° to 175° (fig. 7).

A synoptic diagram (fig. 8) of the data from figures 4, 5, 6, and 7 indicates clearly that β and β'_1 are identical. The identity of the two maxima B and B'_2, although not so convincing, is suggested by similarity of trend and plunge and by the observed relationship of S_2 and B. This is supported by the absence in the field of structures related to directions of folding other than the two obtained statistically. The relationship of the π–S_3 maxima to the other concentrations suggests that S_3 is predominantly a set of 0kl joints associated with each of the two fold axes. On the north coast of Tawharanui Peninsula, although some of the outcrop is obscured by Recent sands, it was seen that part of the succession dips in the opposite sense to the remainder of the strata and is not overturned. This fact, and the variability of attitude, suggests the existence of a fold plunging in a north-north-west direction and

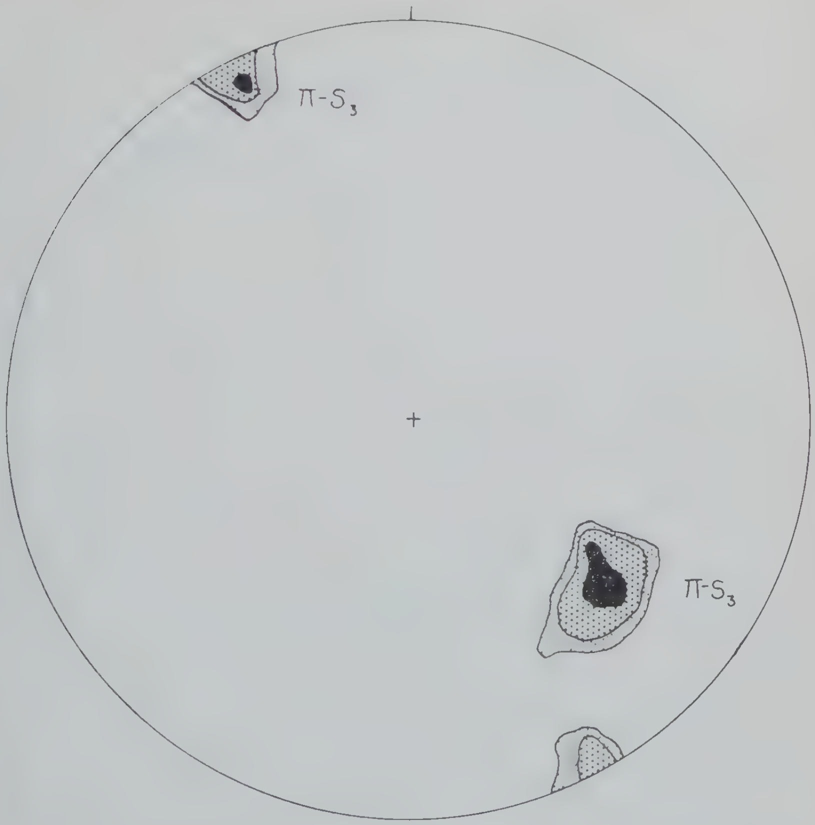

FIG. 6—π-S$_3$ diagram using 154 poles. Contours at 6, 7, 8% per 1% area.

overturned to the north-east. Strata east of the buried hinge dip to both the east and to the west at angles of between 35° and 40° and those on the western limb dip more gently at 10° to 15°. The gently west-dipping strata on the western flank are not inverted but those on the east include both steeply west-dipping overturned strata and uninverted strata dipping gently to the east. Elsewhere on the peninsula, folds too large to be classified as B-corrugations trend to west of north and are thus parallel to the postulated obscured hinge line.

Even with respect to the fold axes alone, the final plot is triclinic and the structure must therefore result from either a triclinic strain or from two independent strains of higher symmetry.

From consideration of the geometry of the structures, and from the styles of folding recognised in the field, the following sequence of events is suggested.

Firstly, flexural slip folding took place about widely spaced β-axes trending at approximately 140° and folds of considerable amplitude and wave-

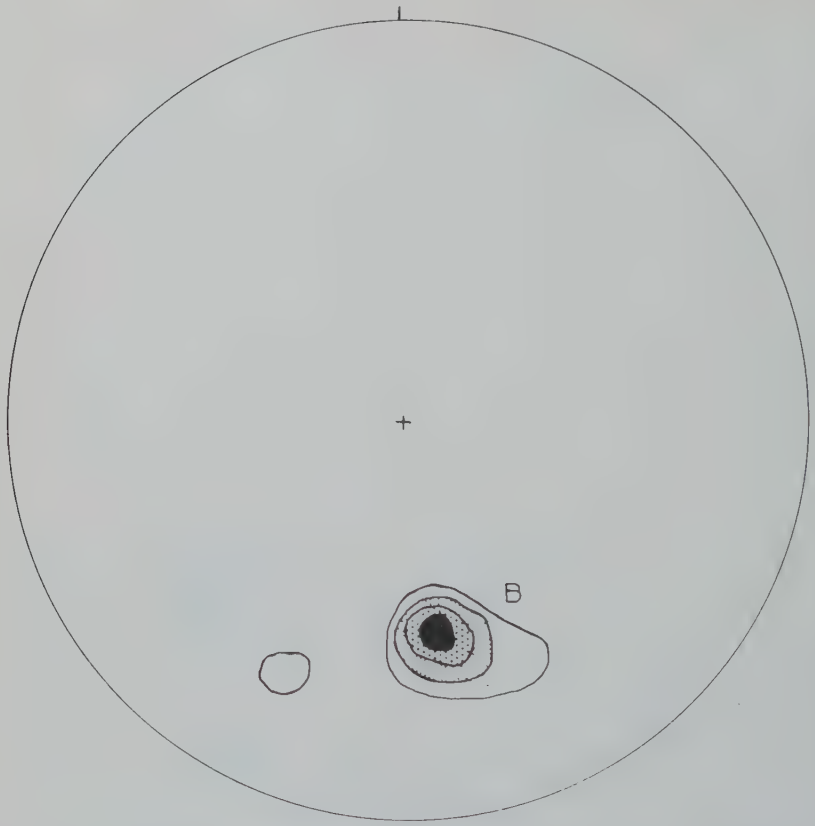

FIG. 7—Plot of B-axes based on 103 poles. Contours at 10, 15, 20, 25% per 1% area.

length were formed. Continued deformation caused steepening of the flanks of these folds until they became isoclinal and were ultimately overturned beyond the vertical by approximately 30° toward the north-east. At a later stage, or in continuation of the earlier phase, after the bedding surfaces became passive consequent on the juxtaposition of the isoclinal fold limbs, flexural slip folding ceased and a set of shear surfaces was initiated inclined at fairly low angles to the direction of principal stress. In this way, the oblique shear surfaces recognisable in the field came into being and movement along them, which was responsible for the corrugations and minor folds parallel to B, absorbed the remaining deformative stresses. Joints due to tension acting perpendicular to the direction of principal stress (De Sitter, 1956, p. 132) were formed during each period of folding. The mechanism of folding which produced B accounts for the absence of curved joint planes.

Although the relationship which the two axes bear to one another was not seen in the field, the styles of folding are compatible with the sequence suggested. Furthermore, the oblique shear surfaces are not markedly affected

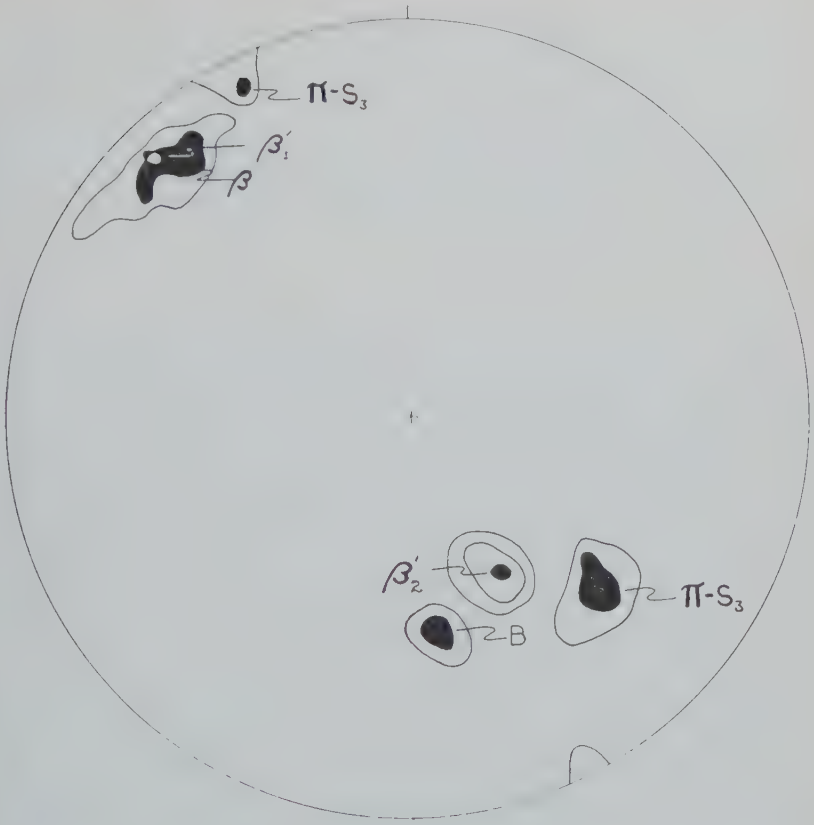

FIG. 8—Synoptic diagram of data from figs. 4, 5, 6, and 7.

by bedding-plane shear but they cut across both massive greywacke sandstones and argillites, and it is thus evident that the first set of folds was produced by flexural slip and that the later folds were thus resulting from movement along inclined shear planes. It is unlikely too, that an initial surface like sub-horizontal planar bedding would remain passive under stress conditions likely to cause the B-axes and then later would become an active surface of slip in response to subsequent orogenic stress. The formation of the small corrugations by oblique shear requires a comparatively high degree of rigidity in the medium undergoing deformation, whereas large overturned folds could form in response to deformation of a much more plastic medium. Hence, most unusual conditions would be required to produce an increase in plasticity of initially indurated rocks (without increasing the superincumbent load so that flow would tend to take place instead of folding), in order to cause a sequence of folding in reverse of that postulated.

The observed lesser degree of deformation in the strata of the eastern end of Tawharanui Peninsula is considered to be due in part to their being sandwiched between thick sandstones. Argillites so placed in an isoclinal fold system might be expected to be shielded from the full effects of deformative stress by the armouring effect of the enclosing sandstone beds. The sandstones and enclosed alternating argillites and sandstones could be expected to respond as a unit so that only the large surfaces of shear cutting the sandstones would be fully effective, and only some of the stress would be transmitted to the central unit. The slight oblique shearing produced along the few S_2 planes so formed would then account for the formation of the observed shallow folds in the very slightly sheared strata.

Conclusions

It is tentatively suggested that the observed pattern of folding may have arisen through the action of compressive forces whose principal stress direction acted along a sub-horizontal direction trending approximately east-north-east. A couple causing overturning towards the north-east must have been active in at least the earlier stages. With overturning of the isoclinal fold system, some clockwise rotation probably took place in the horizontal plane so that earlier formed axial elements in the rocks made an angle of 60° with the principal stress direction which appears to have remained unchanged.

It is clear from the foregoing that, although there is good evidence for a definite pattern of folding in the basement rocks, there is room for more structural work of this type. It seems likely that if collection of a greater number of measurements from much smaller sectors were possible, interpretation of the results so obtained might reveal a much more complex history than that postulated here. It is probable, too, that the regularity of the original structures has been greatly reduced by faults whose displacements it is difficult, or even impossible, to assess because of the lack of distinctive marker beds. The existence of such faulting may be learned only from consideration of the geometry of statistically determined structural elements plotted for small areas.

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POST-GLACIAL TERRACES NEAR CAPE CHOCOLATE, McMURDO SOUND, ANTARCTICA

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Summary

Near Cape Chocolate, two streams fed by the Hobbs and Salmon* Glaciers flow to a bay partly enclosed by a breached loop of an ice-cored terminal moraine of Hobbs Glacier. A bay-head delta constructed by the streams is cliffed at its outer edge, and covers a marine bench which rises to a height of about 30 ft. Along the streams, between the delta and glaciers, are terraces cut in moraine and fluvioglacial sediments.

The sequence provides a basis for a post-glacial chronology of the McMurdo Sound region, but the interpretation is complicated by the interplay of isostatic upwarping following deglaciation and perhaps by eustatic sea-level changes.

A local chronology is presented and the 30 ft bench is tentatively correlated with the climax of the post-glacial rise of sea-level. The present height of the bench is probably due to isostatic uplift.

INTRODUCTION

Cape Chocolate is situated on the west side of McMurdo Sound at the southern-most extent of the sea-ice (fig. 1), and is readily accessible by air from the United States and New Zealand bases at Hut Point and Pram Point respectively, on Ross Island. Possibly the first party to visit the area consisted of Lieut. A. B. Armitage, Dr E. A. Wilson, and Seaman W. L. Heald, members of Scott's Discovery Expedition of 1901–4, who at the end of November 1902 "examined the Koettlitz Glacier" (Scott, 1905, p. 225). They did not specifically mention Cape Chocolate. Subsequently, in February 1911 (Taylor, 1914, p. 558), the Western Party of Scott's "Terra Nova" Expedition, 1910–1913, established a depot on the cape and examined the Salmon Valley (fig. 2) for some 10 miles inland. The cape was named because of the chocolate colour of its moraine (see Helm, 1958, p. 31). The geologists accompanying the Western Party described the area in several publications, the most significant being those by Taylor (1914) and Debenham (1921). Since Operation Deepfreeze commenced in 1955–56, transport facilities provided by the U.S. authorities at Williams Air Facility, Hut Point, have enabled U.S. personnel to visit Cape Chocolate on several occasions (e.g., see Péwé, 1958, p. 1). Similarly, through the courtesy of the U.S. Navy VX-6 Air Command, McMurdo Sound, Messrs. E. B. Fitz-

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*Because of duplication, Salmon Glacier is a new name proposed for the Davis Glacier of Taylor (1911), and others, by Mr G. Warren, Trans-Antarctic Expedition. The name has yet to be approved by the N.Z. Honorary Geographic Board. As a consequence of the change in name, the name Salmon Creek has been proposed instead of David Creek for the stream draining from Salmon Glacier. Hobbs Stream has been proposed for the meltwater stream draining from the Hobbs Glacier.

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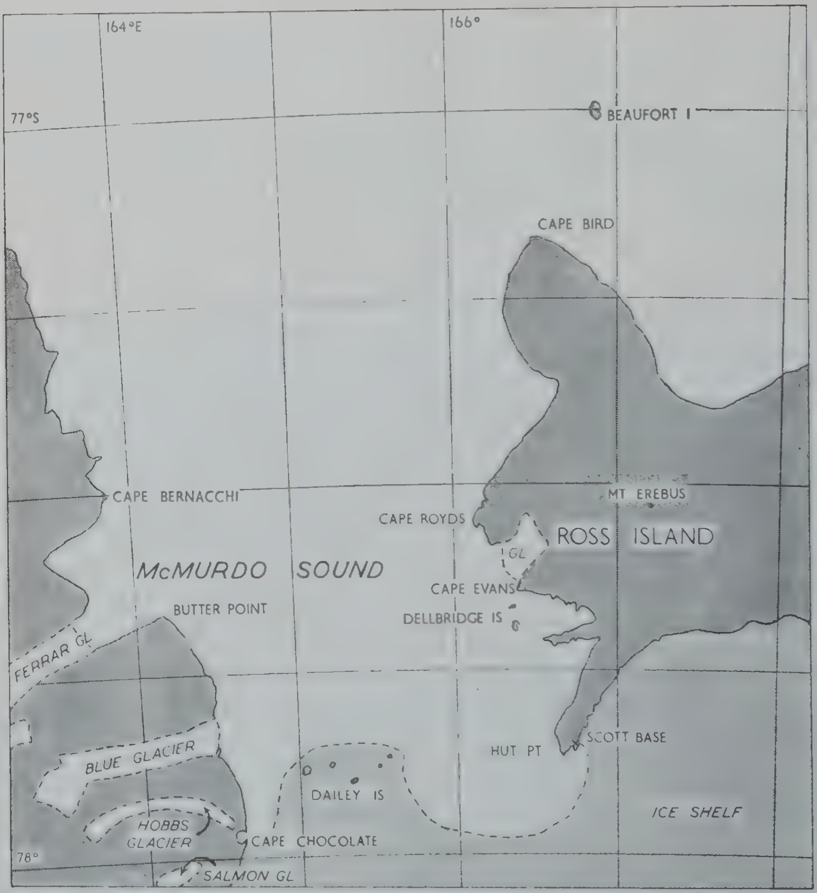


FIG. 1—Locality map showing position of Cape Chocolate in relation to the south end of McMurdo Sound Based on N.Z.M.S. 135.

gerald, J. Wilson, and the writer, members of the 1958–59 New Zealand Geological and Survey Antarctic Expedition, were flown to Cape Chocolate on 28 December, 1958. While a survey station was being established the author had the opportunity to make a quick reconnaissance of the area north of the cape for about 2 miles and inland for half that distance (figs. 2, 8).

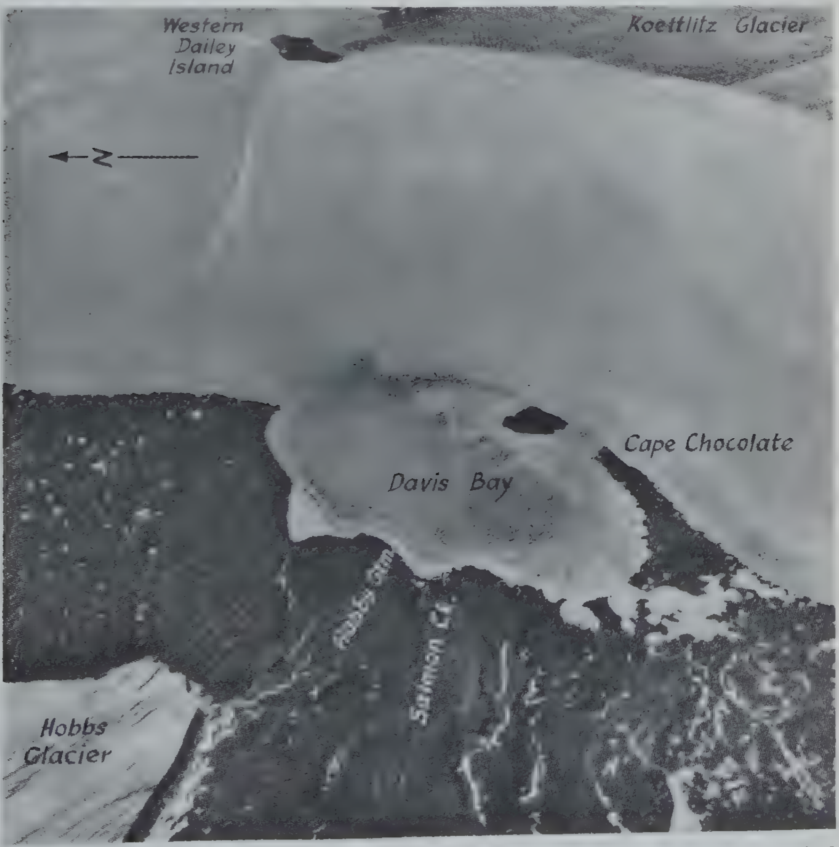
GEOLOGY

Basement Rocks

The foothills of the Royal Society Range behind Cape Chocolate are known from the work of Ferrar (1907), Prior (1907), and Smith and Debenham (1921), to consist of a metamorphic complex of crystalline limestones, marbles, schists and reddish gneissic granites, intruded by lamprophyre dykes. Limestone and marble dominate near the coast, but granite is more common inland.

Quaternary Moraines

A belt of moraine (fig. 3) about 2 miles wide extends from the shore-line up to an altitude of approximately 400 ft and represents lateral moraine deposited during the advance and later ablation of the Koettlitz Glacier (Debenham, 1921, pp. 88–89). In the vicinity of Hobbs and Salmon valleys the lateral moraine of the Koettlitz Glacier has been modified both by the action of, and by the moraine deposited by, Hobbs and Salmon Glaciers. At the head of Davis Bay (figs. 2, 8) the moraine belt is breached by three narrow valleys, the two northern ones being occupied by the seasonal Salmon



Part of U.S. Aerial photo 00095, 6.12.56.

FIG. 2—Oblique aerial photograph of Cape Chocolate region, showing Davis Bay, the bay-head delta and terraces and the meltwater streams.

Creek and Hobbs Stream which flow from glaciers five and one mile inland respectively. The third and southern valley represents a previous course of Salmon Creek and at present is occupied by a very small ephemeral stream.

Little detailed work was attempted on the composition of the moraine. Both the lateral moraine of the Koettlitz Glacier and those of the Hobbs and Salmon glaciers appear to be similar. The moraine is composed dominantly of volcanic material of a wide size range, but also contains about 10% of pebbles and boulders of granite, gneiss, schists, marble or limestone, kenyte and Beacon Sandstone fragments of many lithological types. A fine dark-grey silt forms the matrix of the larger morainic debris and when unfrozen it is easily removed by wind action. Some of the larger boulders are striated. In the vicinity of the contact between moraine and basement, light coloured scree derived from the heavily glaciated, well-rounded foothills, by downward creep during freeze-and-thaw processes, has mixed with the moraine, thus diluting its dark colour (fig. 3).

Cape Chocolate (fig. 4) is cuspate in shape, about 400 yd wide at the base and 100 yd wide at the tip, and extends from the mainland in a northerly direction for about half a mile. Except for a narrow isthmus near the south side it is separated from the mainland by a minor dendritic tidal bay that extends from the south-west corner of Davis Bay. It is composed of moraine heaps of maximum height 60 ft of similar composition to the moraine on the mainland and which probably overlies stagnant ice (Taylor, 1914, fig. 10). The moraine is arranged in marked linear ridges in a north-westerly direction parallel to the axis of the Hobbs Glacier. The largest boulders are mainly concentrated on top of the linear ridges. Approximately 150 yd north of the cape there is a morainic island. The shape of the cape, the orientation of the linear ridges and the present position of the snout of Hobbs Glacier suggests that Cape Chocolate is a remnant of a terminal moraine of Hobbs Glacier.

TABLE 1—Heights and Possible Correlations of the Terraces of Salmon Creek and Hobbs Stream

Hobbs Stream		Salmon Creek	
North Side	South Side	South Side	North Side
9–10 ft to 30 ft (+) bay-head bench and delta			
40–?		28–35	Not measured
55–65	70–?	55–70	Glacial lake deposits under gravels
80–?	80–?		
		110–160 (+)	
One higher terrace level		At least three other higher terrace levels	

Terraces and Delta

Along the sides of Salmon Creek and Hobbs Stream there is a well-developed sequence of meltwater stream terraces. At the head of Davis Bay the fans of the two streams have formed a bay-head delta. The distribution and characteristics of the terraces and bay-head delta are shown in figs. 2–10 and the terrace heights correlated in table 1.

Because of the white and reddish coloured granitic pebbles composing the gravels of the terraces and bay-head delta, they stand out prominently against the darker moraine. The granite pebbles are angular to subangular and very uniform in size, the average maximum dimension being about one inch and seldom greater than three inches. The gravels of the present stream flood-plains are similar except for a higher proportion of larger boulders. Frost polygons are developed on all terrace treads (fig. 7), on all but the steepest terrace faces (fig. 7), on the bay-head delta surface but not on the present stream flood-plains.

The terraces on the south side of Salmon Creek are best developed and cover the largest area (figs. 2, 5, 6, 7, 8). Here there are terrace treads at 28–35 ft, 55–70 ft and the most prominent (figs. 5, 6) at 110 ft to at least 160 ft. (heights given from seaward to inland margins). The last surface continues to rise further inland. At the 160 ft level (fig. 10) the terrace is cut in moraine and lacks a granitic gravel veneer. Downstream from the unveneered portion of the terrace tread there is a break in the continuity of the gravel veneered tread, and the terrace tread cut in moraine at 160 ft may be a slightly higher (no more than 5 ft) level.

Inland, but not measured, are at least three further terrace levels (figs. 3, 10) with successive steps of approximately 30, 10, and 50 ft above the 110–160 -ft terrace surface. On the north side of Salmon Creek the 110–160-ft terrace tread is matched, but the lower terraces have been eroded, leaving only a spur-like remnant projecting into the bay-head delta (figs. 5, 6, 8).

A similar set of terraces has been formed along Hobbs Stream (figs. 2, 8). Only remnants of the terrace levels abutting against moraine are present on the south-west side of the stream, but large areas remain on the north side. When the distance inland is considered, the heights of terraces developed along both streams are comparable (table 1). The 110–160-ft terrace has not been formed. Terrace treads may have minor steps of up to 5 ft.

Fluvioglacial silts (see below) outcrop just below the tread of the terraces (fig. 12) and further upstream moraine is present at the base of faces bordering the flood-plain of Salmon Creek. The terraces are therefore cut in moraine or frozen glacial lake deposits and covered by a thin veneer of gravel.

The bay-head delta is formed by the coalescence of the fans of Salmon Creek and Hobbs Stream. The fan of Hobbs Stream is 2–3 ft higher than that of Salmon Creek so that there is a distinct linear depression at their junction (figs. 5, 8). The delta is cliffed at the seaward edge where its upper surface is 9–10 ft above high tide level (fig. 9). Patches of storm-wave beach occur at the foot of the delta cliffs (fig. 9). Five moraine knolls project above the delta surface, the highest to about 40 ft (figs. 5, 6, 8).

The distribution and form of the terraces differs from that shown by Debenham (1921) on his sketch-map of Davis Valley (fig. 7, p. 74) and sketch (fig. 8, p. 75) of the valley mouth, which are clearly idealised.

Fluvioglacial Sediments

In the terrace face on the north side of Salmon Creek, for about 300 yd upstream from the commencement of the higher terraces, there are outcrops of glacial lake sediments composed of frozen laminated and bedded silts (figs. 11, 12, 13) containing lenses of pebbles and grit and sporadic large boulders (fig. 13). At the furthest inland outcrops the bedding is parallel and dips towards the coast (fig. 11). Further downstream the sediments are cross-bedded (fig. 12). Once exposed to the atmosphere, freeze-and-thaw processes cause rapid disintegration of outcrops which crumble away, in time becoming covered by gravel scree falling from above. Glacial lake sediments were not found along Hobbs Stream.

ORIGIN OF THE TERRACES

Debenham (1921, pp. 75–76) considered that the terraces were cut by streams through "deposits of water-borne silt and wind-borne gravel" formed in a meltwater-lake dammed by the side of the Koettlitz Glacier. The laminated and bedded silts were undoubtedly deposited in a glacial lake, but there may have been other barriers to dam the meltwater from Salmon and later from Hobbs Glacier.

At the climax of the Last Glaciation the glaciers of the foothills region, including the Salmon and Hobbs, would have coalesced to form an ice piedmont which in turn would have fused with the greatly expanded Koettlitz Glacier, the composite ice-sheet filling McMurdo Sound and extending an unknown distance into the Ross Sea. As the glaciers and ice-sheets wasted away after the climax of the Last Glaciation, because of its smaller névé, Salmon Glacier retreated more rapidly than the Hobbs.

A stage would be reached where Salmon Glacier had retreated inland, and in which case, because of the relatively much larger catchment areas of both the Hobbs and Koettlitz Glaciers, either could have acted as a barrier to meltwater derived from Salmon Glacier. The larger névé of the Koettlitz would favour the possibility of the barrier being the ice-wall of that glacier. Yet the probability that Cape Chocolate is a terminal moraine of Hobbs Glacier, the present position and height (80 ft) of the ice-front of Hobbs Glacier and the absence of higher terraces along Hobbs Stream, possibly suggests that the barrier was the ice-wall of Hobbs Glacier. This implies that the level of Koettlitz Glacier decreased more rapidly than that of Hobbs Glacier, a trend which seems unlikely when the extensive névé of Koettlitz Glacier is considered. However, once seawater commenced to circulate beneath the base of the Koettlitz Glacier it may have melted at a more rapid rate than Hobbs Glacier. Another possible barrier would have been a residual coastal piedmont, similar to those along the west side of McMurdo Sound further north. Whatever the barrier, it was probably continually breached, resulting in a changing base-level and periods of vertical corrasion of meltwater streams. The higher terraces are developed only along Salmon Creek, but the presence of matching sets of the lower terraces along Hobbs Stream and Salmon Creek suggests that the factor or factors controlling the formation of the terraces continued to operate after the meltwater barrier had ceased to exist.

Godwin *et al.* (1958), has presented evidence indicating a rapid eustatic rise of ocean level between 14,000 and 6,000 years ago, to a maximum suggested extent of about 450 ft. The postulated rate of ocean-level rise of

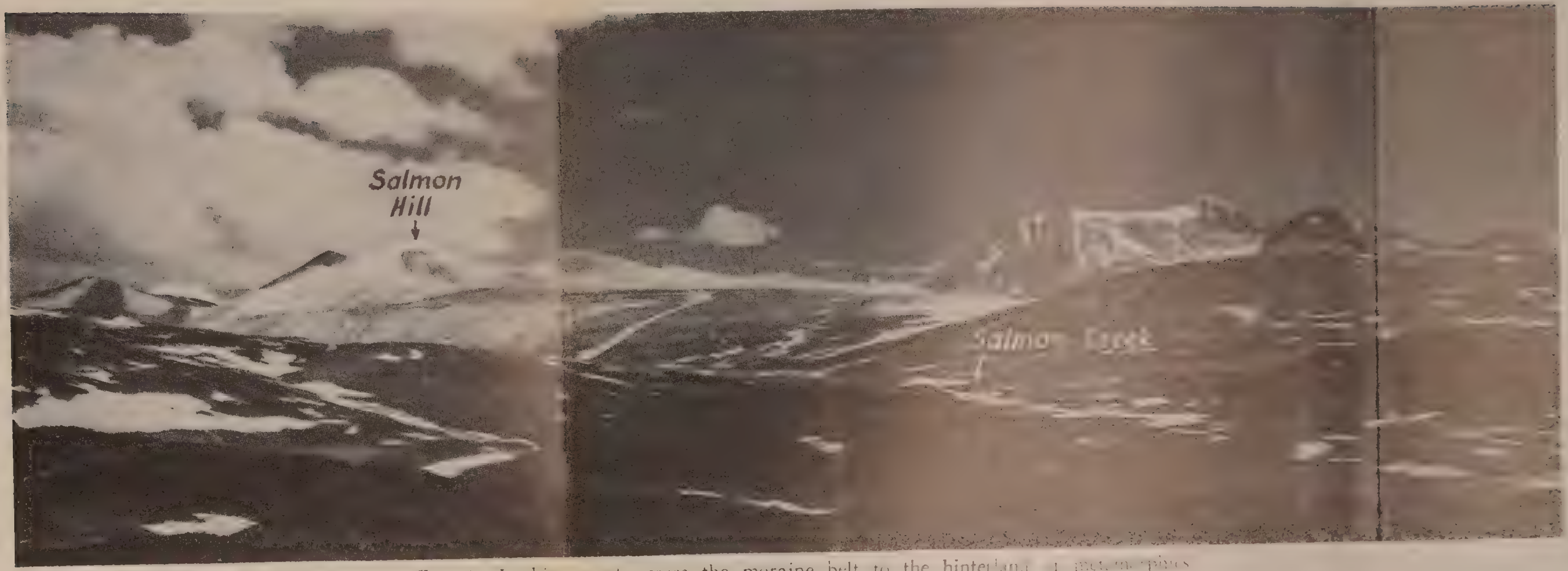


FIG. 3—Looking west across the moraine belt to the hinterland of metamorphics intruded by thin dykes. Salmon Creek flows down a gorge between the hills on the left of the panorama and across the centre. Well-developed terraces veneered with light coloured gravel are shown along the lower reaches of Salmon Creek.

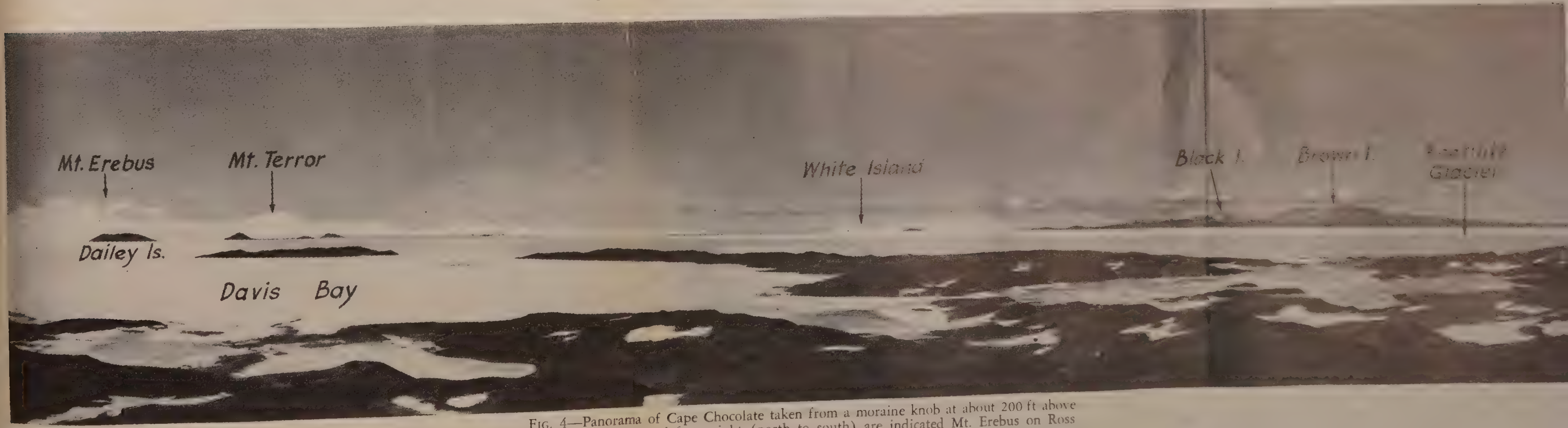


FIG. 4—Panorama of Cape Chocolate taken from a moraine knob at about 200 ft above sea level. From left to right (north to south) are indicated Mt. Erebus on Ross Island, the Dailey Islands, White Island, Black Island, and Brown Island. Note the tidal depression separating the cape from the mainland.

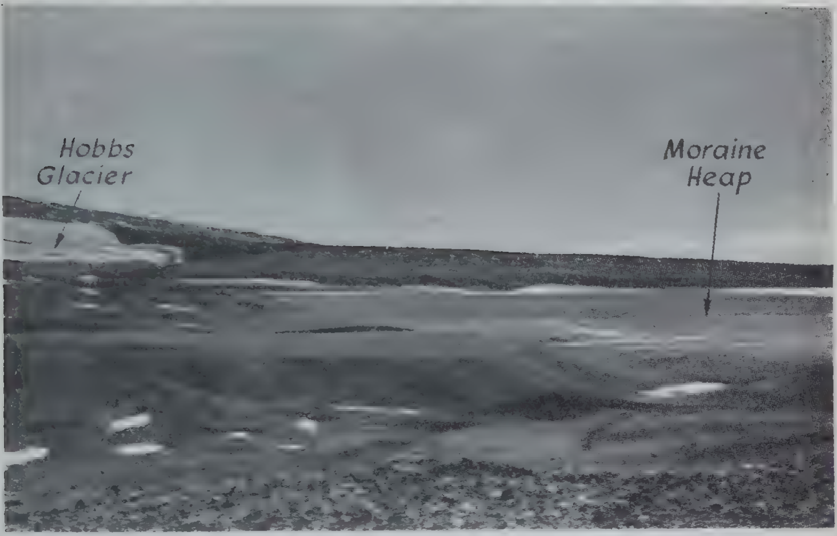


FIG. 5—A view looking north across the bay-head delta to the snout of Hobbs Glacier and the moraine of the north side of Davis Bay. Taken from a moraine knob situated south of Salmon Creek terraces. The terraces on the south side of Salmon Creek are separated from the moraine by a gorge. A linear depression, separating the fans constructed by Salmon Creek and Hobbs Stream, passes across the centre of the photograph and is emphasised by patches of snow and by water derived from Salmon Creek. The cliff at the back margin of the bay-head delta is evident in the middle distance as are the terraces on the north side of Hobbs Stream. Slightly darker moraine mounds projecting through the delta surface are shown between distributaries of Salmon Creek. See fig. 8 for plan of area.

approximately 3 ft per century is less than the estimated rate of isostatic upwarping following deglaciation in previously extensively glaciated regions such as Fennoscandia and Canada Since the Last Glaciation there has been upwarping of approximately 1,700 ft in Fennoscandia and of some 900 ft in Canada, at an average rate of about 12 ft per century for the last 8,800 years in the former region (Flint, 1957). It can be reasonably postulated that isostatic adjustment following deglaciation has also been operative in Antarctica and that it probably had more effect than post-glacial eustatic sea-level changes. Therefore isostatic adjustment was perhaps important in controlling the alteration of base-level resulting in stream downcutting and the formation of terraces.

It is difficult to reconcile the expected continuous regional upwarping with the formation of the 30-ft bench and the cliffing of the delta. Hough (1950) has shown (see below) that climatic fluctuations have occurred in Antarctica during post-glacial time. Therefore, at certain intervals, climatic fluctuations may have resulted in sea-level changes of sufficient significance to cause the formation of the 30-ft beach and the cliffing of the delta.

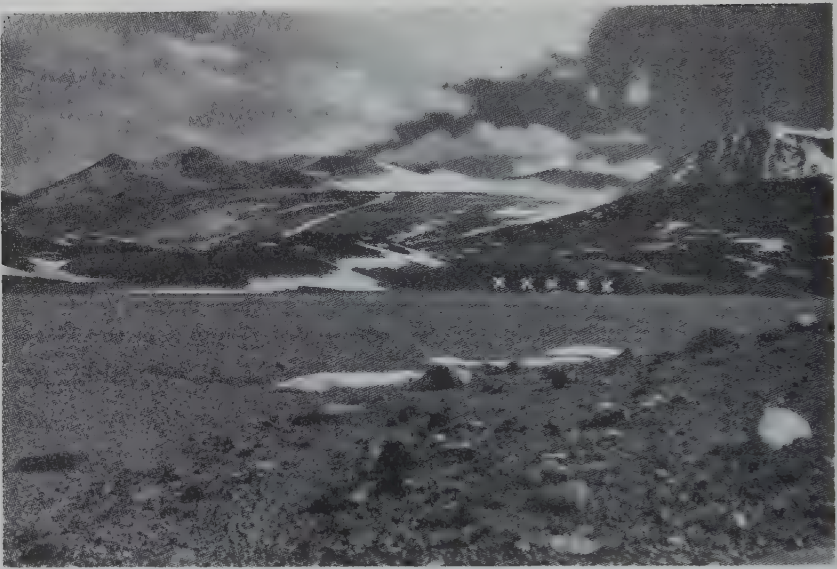


FIG. 6—Photograph of the bay-head delta from its seaward extremity looking inland on to the terraces at the mouth of Salmon Creek. The delta surface has frost polygons, and the lower slope of a moraine heap projecting through the surface is present on the right foreground. The gully separating the terraces on the south side of Salmon Creek from the moraine is apparent in the left middle distance of the photograph. The sequence of terraces on the south and north sides of Salmon Creek are clearly shown. The positions of the frozen glacial silt outcrops are indicated by crosses. Note contact of light coloured gravels and dark moraine on south side of Salmon Creek in middle distance, just above snow patch.

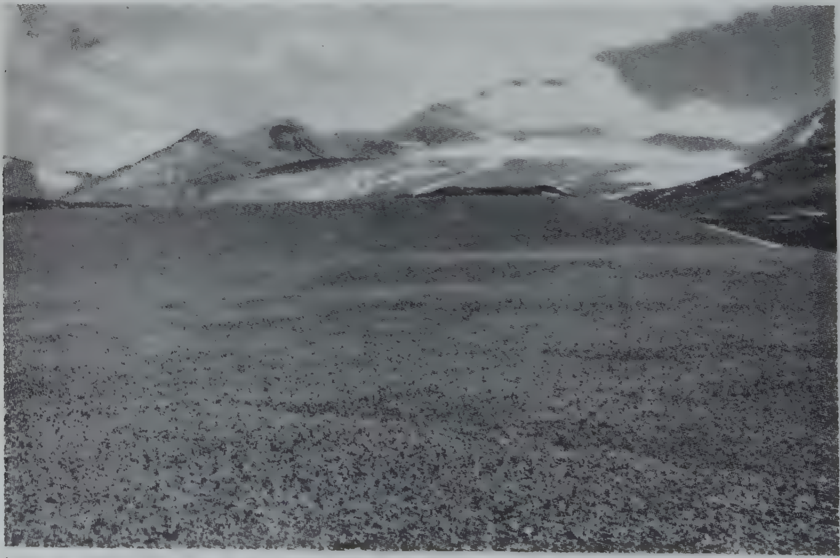
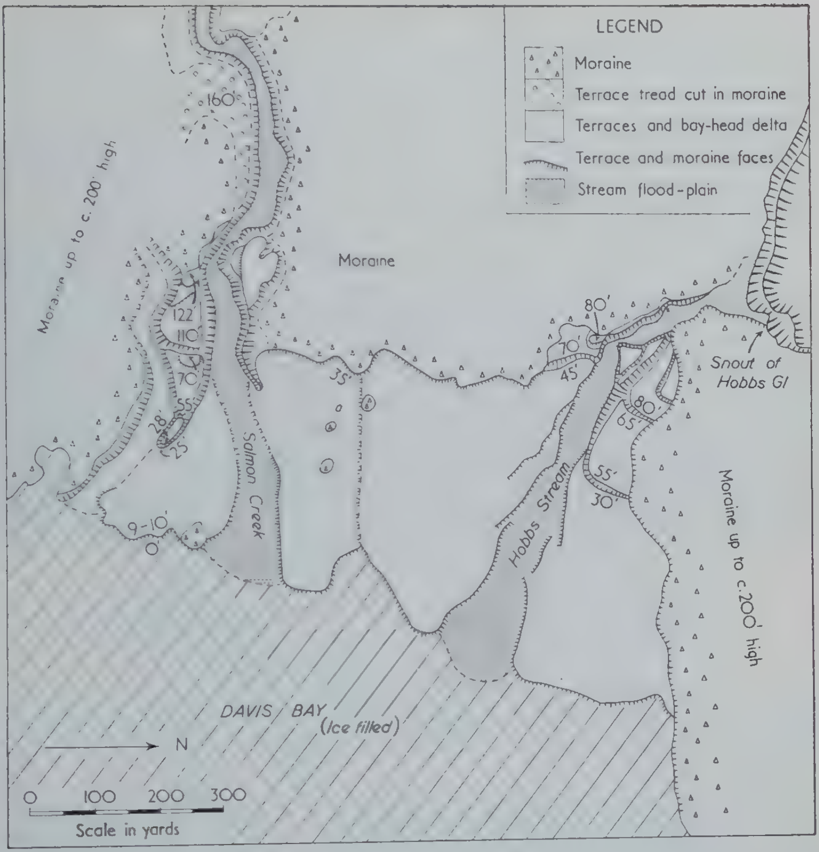


FIG. 7—Frontal face of the 110-160 plus-foot terrace and the tread of the 55–70 ft terrace on the south side of Salmon Creek, showing pebble size, nature of surface, and frost polygons.



Drawn from U.S. aerial photograph 00148.

FIG. 8—Sketch map of the delta and terraces at the head of Davis Bay. The heights and points where heights of terraces were measured are indicated by the position of numerals.

AGE AND CORRELATION

The deposits are clearly post-glacial. At Cape Chocolate the relative ages of the features described above are distinct (table 2). Silts with discontinuous grit lenses and pebbles were deposited in a glacial lake. At some stage the lake was drained and the silts frozen. Terraces were cut in moraine and frozen glacial lake deposits and the seaward parts of terraces and moraine benched. Meltwater streams then constructed fans on the bench to form the present bay-head delta. Subsequently the outer edge of the delta was cliffed. Finally, there has been a relative slight recession of sea-level resulting in small prograded beaches in front of the cliffed delta margin.

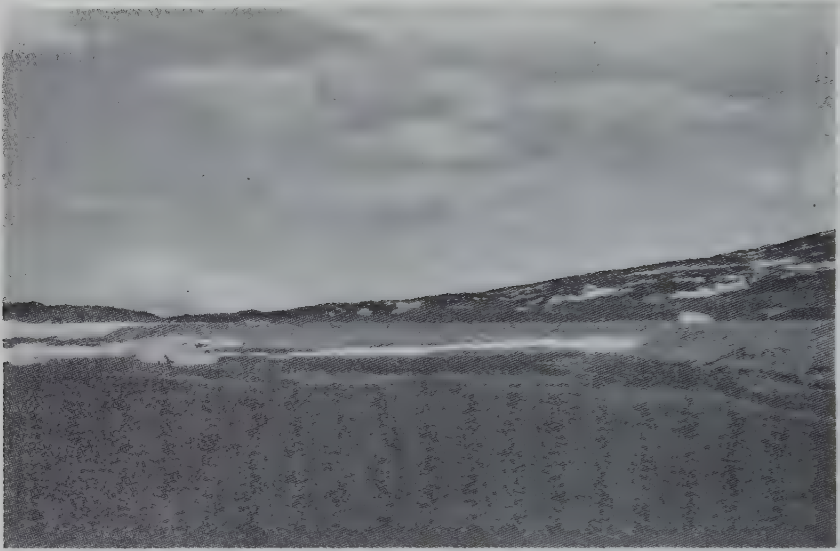


FIG. 9—Photograph looking south towards the head of the Koettlitz Glacier and showing the cliffed, here 9 ft high, outer margin of the bay-head delta. An area of "storm-wave" beach is here present in front of the cliff. The high tide mark is indicated by the lowest extent of the light-coloured dried algae and salt encrustations.

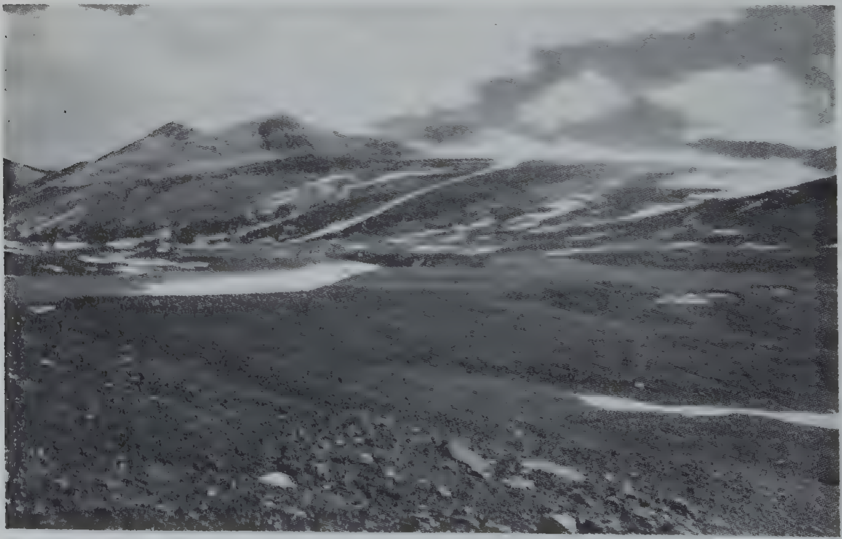


FIG. 10—A view looking up Salmon Creek which is entrenched between terraces near the centre middle distance of the photograph. At the downstream extent of the snow patch, which is on the south side of the creek, the creek turns sharply to flow north-east beyond the right margin of the photograph. The level surface in the foreground is terraced moraine lacking a gravel veneer. Inland further flights of terraces are visible.

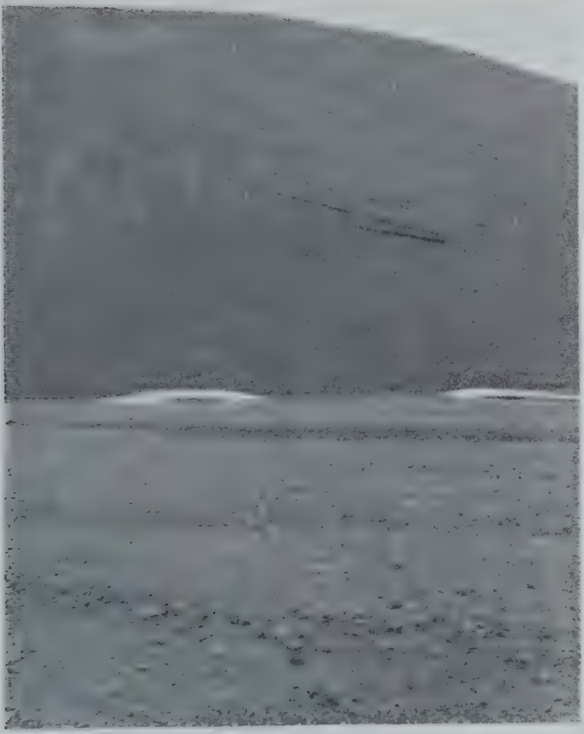


FIG. 11—The face of a terrace, the top of which is about 30 to 35 ft above the present flood-plain of Salmon Creek, showing bedded and laminated glacial lake deposits dipping downstream.

Hough (1950) has described the alternation of glacial marine and marine sediments in cores taken from near Scott Island, north of the Ross Sea. Following a cold period centred about 21,000 years ago and extending to approximately 15,000 years ago, there was a warmer period which lasted to about 6,000 years ago. The warmer period was then followed by a colder period which continues to the present day. This suggests that the cold conditions ameliorated somewhat earlier in Antarctica than in other parts of the world (10,000 years ago, Harrington, 1958), and that the terraces and glacial lake silts are younger than 16,000 years ago. Ablation must have taken a considerable time, so that there will be an unknown interval before the deposition of the fluviatile deposits. This is certainly true as the southern-most core described by Hough is some 15° north of Cape Chocolate.

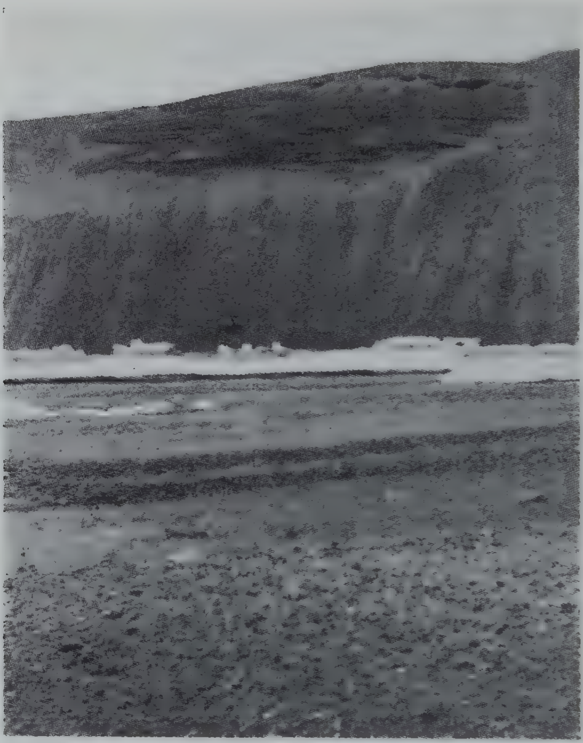


FIG. 12—Cross-bedded glacial lake deposits underlying a terrace surface on the north side of Davis Creek. One branch of Salmon Creek is present in the foreground. The small Bergen pack on the face of the terrace gives an indication of scale.

TABLE 2—Chronology of the Cape Chocolate Sequence and Possible Correlatives in the McMurdo Sound Region

Sequence at Cape Chocolate	Correlatives in McMurdo Sound Region
Progradation of present shore	Progradation at Cape Bird
Cliffing of delta	Cliffed stream fans at Cape Bird and talus at Franklin Island
Building of bay-head delta by meltwater stream fans	Meltwater stream fans at Cape Bird
Cutting of 30 ft + marine bench	25 ft bench at Beaufort Island, Franklin Island, etc.
Down-cutting of Salmon Creek and Hobbs Stream	
Deposition of fluvioglacial beds	? Similar features in neighbouring valleys
Ponding of Salmon Creek and Hobbs Stream to at least *c.* 70 ft above present sea level	

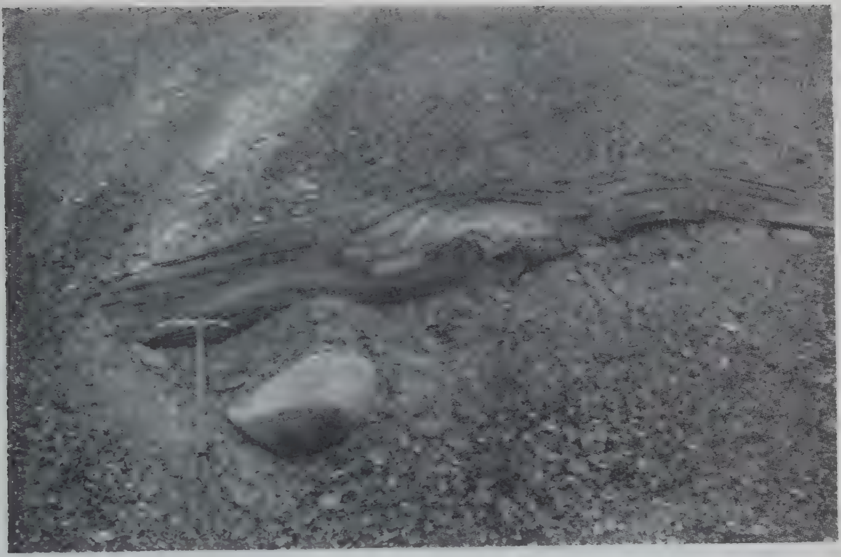


FIG. 13—Close-up photograph of a bedded and laminated glacial lake silt outcrop with an enclosed rounded granitic boulder. The head of the ice-axe is 9 in. long.

A possible datum plane in the chronology of the Cape Chocolate sequence is the 20–30-ft bench which is apparently widespread around McMurdo Sound and beyond. Similar benches have been recorded at Cape Bernacchi (20 ft and 40 ft; David and Priestley, 1914, p. 84), Taylor Dry Valley (20 ft plus; David and Priestley, 1914, p. 95), Cape Adare, Wood Bay, and Franklin Island (Ferrar, 1907, p. 81), Beaufort Island (25 ft; Harrington, 1958, p. 602), Franklin Island (Brodie, 1959, pp. 114–5) and Granite Harbour (30 ft; A. C. Beck, pers. comm.). Like the Cape Chocolate bay-head delta, the benched talus apron at Franklin Island is fronted by low (10 ft) cliffs at the shoreline.

For the cliffed coastal bench at Beaufort Island, if marine, Harrington (1958, p. 602) suggested formation during the Climatic Optimum, about 5,000 B.C. The raised bench on Franklin Island was thought by Brodie (1959) to be probably related to the raised sea level of the Climatic Optimum. According to Suggate (1958, p. 121) the post-glacial rise of sea-level in New Zealand, at Christchurch, reached a little above its present height about 5,000 years ago. The McMurdo Sound benches may be related to the culmination of this post-glacial rise of sea-level and be of comparable age. Hough's (1950) study suggests that the bench was formed prior to 6,000 years ago. Post-glacial benches in other regions are apparently less than 10 ft high (Godwin, 1956), and as suggested by Harrington (1958, p. 602) those around McMurdo Sound were probably raised to their present position by isostatic uplift.

There is little support for eustatic sea-level fluctuations related to the climatic changes indicated by floral successions in Europe and elsewhere (Godwin, 1956). Harrington and McKellar's (1958, p. 573) study of the Cape Hallett Adelie penguin colony suggests that there has been little climatic fluctuation in the last 1,200 years, which is in agreement with Hough's (1950) work, although the marine sediments may not necessarily record minor climatic fluctuations. If minor climatic fluctuations controlled the cliffing of the delta and subsequent relative retreat of sea-level at Cape Chocolate, then the fluctuations probably occurred before the last 1,200 years.

Suggested correlations of the Cape Chocolate sequence to similar features observed by Dr H. J. Harrington, N.Z. Geological Survey (now at University of New England, Armidale, Australia) and Dr C. Johnson, U.S. Geological Survey, at Cape Bird and Beaufort Island, are given in table 2.

COMMENTS

The foregoing is a description of the post-glacial sequence developed in one of the several "dry" valleys cutting the foothills to the east of the Royal Society Range, on the west side of McMurdo Sound. Photographs of the coastal strip, from Heald Island northwards, show comparable meltwater stream deposits in other "dry" valleys, particularly in the Miers and Garwood valleys. It is hoped that this paper draws attention to an area that may hold the key to the late Pleistocene chronology of the McMurdo Sound area – certainly of the post-glacial chronology.

ACKNOWLEDGMENTS

The author is indebted to the U.S. Navy Air Squadron VX-6 at Williams Air Facility for transport to Cape Chocolate. Thanks are extended to other members of the 1958–59 N.Z. Geological and Survey Antarctic Expedition for their assistance in the field and for discussions; and especially to Drs H. J. Harrington and C. G. Johnson for permission to publish some of their data from Cape Bond, Ross Island.

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HUTT VALLEY FAULT SCARPS

By C. A. Cotton, Lower Hutt

(Received for publication, 14 October 1959)

Summary

It is improbable that the sinuosity of the Wellington Fault scarp has been notably increased by localised back-wearing. North-east of the Wakatikei River the valley-side scarp of the Hutt Valley is half a mile out of alignment with the Wellington Fault and it is possible that its position is determined by a splinter or *en échelon* fault. The long straight segment of the valley side south-west of the Wakatikei confluence has not necessarily been cut far back by the Hutt River, which more probably has merely retouched the youthful form of a fault scarp, perhaps removing soliflual debris that threatened to smother it. The view that the valley-side scarp is a young fault scarp is supported by the asymmetric form of the valley.

Comparative immunity of the right bank of the Hutt River from attack by lateral river corrasion for some distance below the Akatarawa confluence may have been due to fan-building by the Akatarawa River.

Offsets in the scarp of the Wellington Fault have been attributed by Stevens (1958) in one case (south-west of Korokoro) to an embayment due to retrogradation by the action of waves on Port Nicholson and in another case (bordering the Hutt River in the Upper Hutt basin) to lateral river corrasion. There can be no doubt that erosion by harbour waves has caused some retreat of the scarp, but the extent of this has been over-estimated by both Hall and Stevens (see Cotton, 1957, p. 773). Lateral corrasion by the Hutt River must also have taken place, freshening the facets of the scarp. Despite discovery of outcrops of bedrock in the river bed in front of the scarp, however, the extent of the erosional retreat of the valley-side scarp from the initial fault scarp in the lower valley may very well have been much less than the 400 or 500 yd Stevens (1957, fig. 15) has shown in a section or even the 200 or 300 yd he has more recently assumed (1958, p. 643). This estimate has been based on the distance from the scarp at which bedrock was found at a depth of about 100 ft in two bore holes. The presence of bedrock at shallow depth in front of the scarp can be accounted for by step faulting, and Stevens (1958, p. 644) has himself expressed belief in "an anastomosing pattern of minor faults" in the Wellington Fault zone.

Attempting to explain the offset in the valley-side scarp at the Wakatikei River confluence (fig. 1), Stevens (1958, p. 642) has written: "It appears that the scarp south of the confluence is largely erosional [meaning that it has been cut far back from the initial form]; the fault scarp to the north has undergone comparatively little retreat from the original fault line." He has assumed here that the initial forms of these two segments of scarp were collinear on a straight Wellington Fault trace. If 200 or 300 yd is an excessive estimate of erosional retreat of the scarp in the lower valley, the half-mile or more that must be postulated to explain this offset of the valley-side scarp as due to erosion is out of the question.

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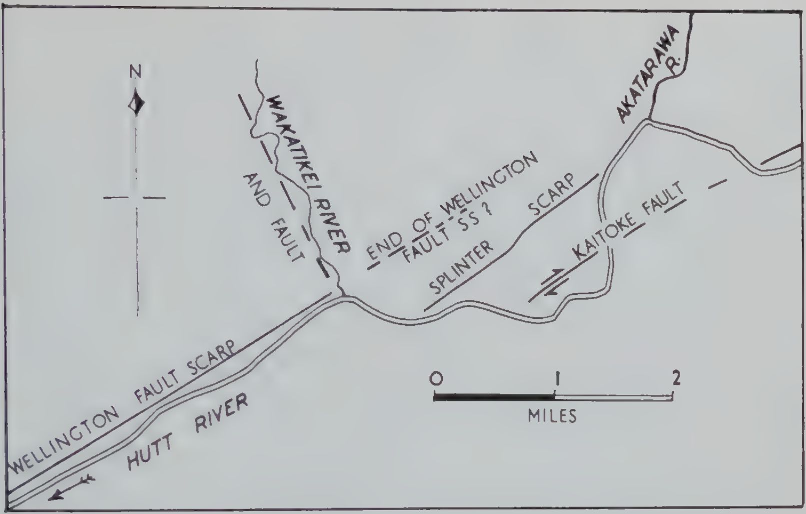

FIG. 1—Faults *én echelon* in the Hutt Valley.

If it be granted that the offset in scarp alignment is not caused by differential erosion, it can be accounted for by postulating either that the Wellington Fault itself is offset by a cross fault in the obviously tectonic valley of the Wakatikei or, perhaps more probably, that the scarps were made by faults *en échelon*. Lensen (1958, fig. 7) has assumed that the Kaitoke Fault is an extension *en échelon* of the Wellington Fault which takes up the dominant transcurrent displacement of the latter; and this view has been adopted by Stevens (1958, fig. 2). Stevens (1956, fig. 2) has mapped a fault extending north-eastward from the Wakatikei River almost in a direct line with the valley-side scarp farther down the valley, and it seems quite possible that this is the Wellington Fault *sensu stricto*, which, however, must die out in a short distance beyond the Wakatikei (fig. 1). The fault along the scarp of the trans-Wakatikei offset may thus be an intermediate *en échelon* fault between the Wellington and Kaitoke faults (fig. 1), and if that is the case the offset is a splinter.

The trace of this fault may be at the foot of the "splinter scarp", but possibly it has been buried under solifluxion coulées that have built the scarp forward and are now thoroughly dissected. Rapid dissection is implied by the topographic form, and this is not inconsistent with the presence of an apron or at least a veneer of soliflual debris on the valley-side face of the splinter. The texture of dissection is fine and the spurs between the numerous ravines taper to points (Stevens, 1958, plate 46).

Though it is highly improbable that the offsetting of the valley-side scarp is due to any great extent, or entirely as Stevens has supposed, to differential erosion, his description of the trans-Wakatikei offset, or splinter scarp, as immune from undercutting by the Hutt River for a very long time is undoubtedly correct. Though it is bordered by a flight of terraces (Stevens, 1956, fig. 2) which indicate occasional swinging of the river to this side of the valley, this scarp has not been recently undercut, and the contrast in topographic form between it and the undercut valley-side downstream from the Wakatikei confluence is real and very striking.

Under the heading "prefrontal solifluxion coulées" Meynier (1958, p. 394) has discussed a somewhat similar case in Brittany. The south wall of the Rennes basin, which elsewhere slopes only gently, becomes a sharply defined scarp at each side of the opening of the gorge by way of which the Vilaine River leaves the basin. The local sharpening of the weak slope into a scarp had earlier been attributed to lateral corrasion by the Vilaine, but Meynier has now pointed out that, though that river has no doubt swung laterally and kept the escarpment cleanly swept, this has not involved erosion of solid rock. The slope northward into the Rennes basin is mantled with solifluxion coulées and these reduce its steepness except in the vicinity of the river, which has eroded the soliflual debris and carried it away. The Hutt River may in this way have kept the Wellington Fault scarp farther down the valley freshly trimmed and so have preserved its youthful form without cutting it back appreciably. The river could from time to time undercut the scarp, restoring its sharpness by removing accumulated soliflual debris.

If the theory were seriously entertained that the scarp along the north-west side of the Hutt Valley was due largely to erosion, instead of being a young fault scarp, it would no longer be possible to explain the very striking asymmetry of the form of the valley as due to its fault-angle origin. Notwithstanding the fact that the Hutt River happens at present to hug the north-west side of the valley for many miles there is nothing to prevent it from swinging widely on its valley floor, and indeed it must have done so in the past in order to distribute the alluvial gravel with which the valley has been infilled. If the river had cut far back and had thus developed a line of erosional bluffs along the north-west side of the valley it would, therefore, be hard to understand its failure to undercut the other side of the valley as well and to explain why the deeply weathered spurs between embayments of the south-east side have not been truncated or even cut away altogether so as to destroy the tectonic asymmetry of the valley.

One question that has not been satisfactorily answered is why the Hutt River, though it now hugs the right bank for the greater part of the length of the valley and probably has undercut this bank from time to time in the past, has yet tended to swing away towards the left for probably some tens of thousands of years so as to leave the "splinter scarp" immune for most of the time from attack by lateral corrasion. The explanation of this tendency given by Stevens (1958, p. 642), namely, that a jet of inflowing water channeled through the gorge of the Akatarawa tributary has in the past thrust the main river across the valley, and still does so, is not acceptable, for it is not the volume of water in a tributary stream or the velocity of its flow but the solid load it brings into the main valley that may deflect the main stream. The Akatarawa could do this, however, and perhaps has done so for long periods in the past by maintaining an alluvial fan across the Hutt Valley which the Hutt River has lacked the power to cut away.

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OUTLINE OF THE CRETACEO-TERTIARY SEDIMENTATION IN THE EASTERN BASIN OF NEW ZEALAND

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(Received for publication, 2 October 1959)

Summary

Correlation by lithology of stages in the Eastern Basin of New Zealand is outlined. Lateral facies changes in the Eastern Basin are consistent in time and place from the Upper Cretaceous to the Lower Tertiary and indicate that finer sediments were laid down in the central, presumably deepest part of the basin. After the close of the Eocene relief in the Eastern Basin became irregular and a sedimentation inconsistent in place and time took place. Tectonic deformation affecting the Eastern Basin was apparently not accompanied by severe breaking up of the undermass prior to the Middle Cretaceous, but was so in the time afterwards. The central area of the Eastern Basin and the relief of Central New Zealand appears to have been tectonically low during the Upper Cretaceous and Lower Tertiary. After the Eocene the central area of the basin broke up into an irregular arrangement of tectonic highs and lows while the relief of Central New Zealand remained low.

INTRODUCTION

The strata of the Eastern Basin (Kingma, 1957a) may be followed from the Raukumara Peninsula through Hawke's Bay and Marlborough into North Canterbury. The lithological units and the regional facies changes that have been observed are, particularly in the Upper Cretaceous and Lower Tertiary, so consistent in place and time and so inconsistent afterwards that important conclusions as to the regional deformation of the Eastern Basin can be drawn from them. Extensive micropaleontological work carried out by the Micropaleontological Section of the New Zealand Geological Survey over the past fifteen years has moreover made possible a reliable time correlation of the great diversity of strata throughout the Eastern Basin. Apart from personal observations in the field and from examinations of sample residues in the Geological Survey collections, data have been derived from the references listed and from Mr N. de B. Hornibrook's description of the Upper Cretaceous and Tertiary stages in the Stratigraphic Lexicon for New Zealand (C. A. Fleming, Ed. (1959)).

The correlation of the strata and facies changes will be discussed stage by stage from the Haumurian onward. The stages and their tentative equivalents, after Hornibrook (1958), are listed in the following table:

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New Zealand Geological Survey		European Equivalent		
Series	Stage	Stage	Period	
Wanganui	Castlecliffian	Sicilian		
	Nukumaruan	Calabrian	Pleistocene	
	Waitotaran	Astian- Plaiscancian		
	Opoitian		Pliocene	
Taranaki	Kapitean	Pontian	(Upper)	
	Tongaporutuan	Sarmatian		
Southland	Waiauan	Tortonian	(Middle)	Miocene
	Lillburnian	Helvetian		
	Clifdenian	Burdigalian	(Lower)	
	Altonian	Aquitanian		
Pareora	Awamoan			
	Hutchinsonian		(Upper)	
	Otaian	Chattian		
			(Middle)	Oligocene
Landon	Waitakian			
	Duntroonian			
	Whaingaroan	Lattorfian	(Lower)	
Arnold	Runangan	Priabonian	(Upper)	
	Kaiatan			
	Bortonian	Lutetian	(Middle)	Eocene
	Porangan			
Dannevirke	Heretaungan			
	Mangaorapan	Ypresian	(Lower)	
	Waipawan		Paleocene	
Mata	Te Urian	Danian	Cretaceous	
	Haumurian	Maestrichtian		

STRATIGRAPHY

Haumurian Stage

The Haumurian Stage can be traced throughout the Eastern Basin. The lithologies range from argillites, sandstones, with grits in the Raukumara Peninsula (M. C. Pick, pers. comm.; pers. observ., and a reinterpretation of the Cretaceous stratigraphy of Ongley and MacPherson, 1928; Henderson and Ongley, 1920; Ongley, 1930) to grits and sulphur sands along the coasts of Hawke's Bay (Kingma, 1957a), Wairarapa (Lillie, 1953; and pers. observ.), Marlborough (MacPherson, 1949; and pers. observ.), and northern Canterbury (Wilson, in press). The argillaceous rocks in this stage, usually with conspicuous concretionary horizons, better known as the Whangai Shales, form the most striking strata in the Upper Cretaceous and may be traced from minor occurrences in the Raukumara Peninsula into well developed sequences in Hawke's Bay, Wairarapa, and Marlborough (Kingma, 1957a; Lillie, 1953; MacPherson, 1949, 1952; Wellman, 1955), while further south the shale grades into the coarser Haumurian strata of southern

Marlborough and northern Canterbury (Suggate, 1958; Wilson (in press); Wellman, 1955; MacPherson, 1949), where depositional conditions were obviously much more shallow. The diminishing grain size in a westerly direction as has been observed in Hawke's Bay (Kingma, 1957a) and Wairarapa (pers. observ.) is significant information bearing on the outline of the Eastern Basin at the time, also the fact that these facies changes are to be seen only in Hawke's Bay and Wairarapa, and probably also in Marlborough and not elsewhere yields important information as to the distribution of deeper parts, but not necessarily of depths in the Eastern Basin. The general outline is thus that shallow conditions prevailed in the Raukumara Peninsula and along the coasts of Hawke's Bay, Wairarapa, and Marlborough into northern Canterbury generally forming an arc on the concave side of which finer sediments were deposited in Hawke's Bay, Wairarapa, and Marlborough. The author had come earlier to the conclusion (Kingma, 1957a) that west of the Hawke's Bay area the New Zealand geosynclinal conditions lasted for a longer time than elsewhere. These conditions presumably stretched as far south as Marlborough.

Te Urian Stage

As in the previous stage the Te Urian deposits show a similar distribution in fine and coarse grained strata. Consisting mainly of blue siltstones and argillaceous shales, occasionally bentonitic and with carbonaceous horizons, the stage is developed as a greensand overlain by a dark brown sandstone in the eastern part of Hawke's Bay and Wairarapa and as a blue, muddy siltstone overlain by black, muddy shale in the western parts (pers. observ.). The stage has so far not been recognised satisfactorily on microfaunas in Marlborough, but the lowermost part of the Amuri Limestone and flint beds in the Wairarapa belongs to the Te Urian (G. H. Scott, pers. comm.), and as the lower part of the Amuri Limestone in Marlborough belongs to the Dannevirke Series (MacPherson, 1949, 1952; G. J. Lensen, pers. comm.), the Te Urian Stage has to be sought in the base of the Amuri Limestone and in the extensive chert beds immediately underlying it. In North Canterbury sulphurous greensands indicate a shallowing (Wilson, in press) while still further south the stage is represented by terrestrial deposits.

The sediments in these Cretaceous stages thus clearly indicate that in the northern and southern extremes of the Eastern Basin shallow conditions prevailed, particularly in the south. These shallow conditions can be traced through the eastern part of Hawke's Bay, the Wairarapa, and probably Marlborough, but in the western parts of these land districts deeper-water conditions were prevalent. In this central part the sedimentation of the New Zealand Geosyncline still continued, while further north and south the New Zealand Geanticline had already appeared. Thus, during Upper Cretaceous time the Eastern Basin was flanking the New Zealand Geanticline in the Raukumara Peninsula and in the South Island, but in the central part of New Zealand the basin merged into the remnant of the still existing New Zealand Geosyncline (fig. 1A). It is in this depression that the Whangai argillites and the fine grained Te Urian strata were laid down; that the Whangai argillites with their striking concretionary horizons can be found in Marlborough (Wellman, 1955; G. J. Lensen, pers. comm.) empha-

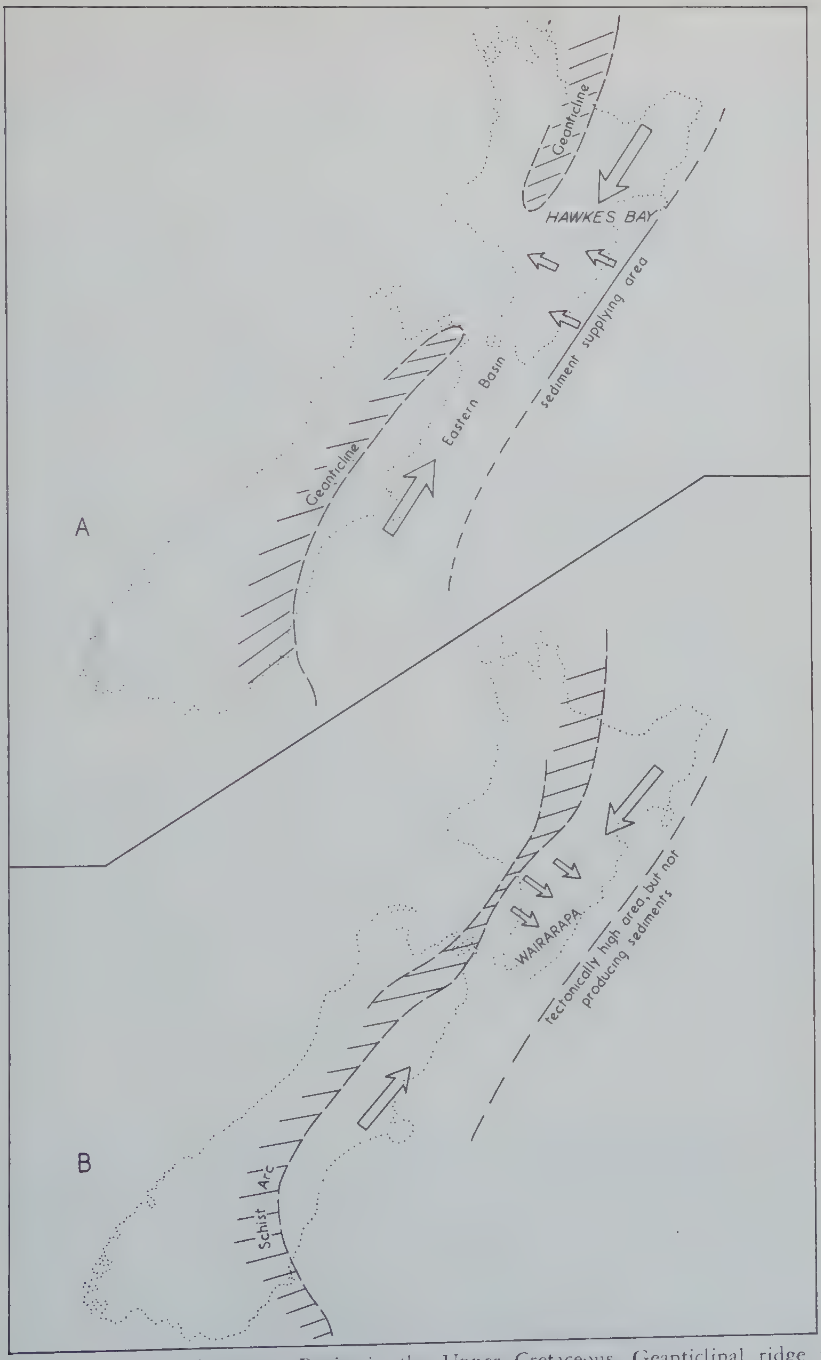

FIG. 1A—Outline of Eastern Basin in the Upper Cretaceous. Geanticlinal ridge is shown and tectonically high area to east of Basin. Hollow arrows indicate direction in which grainsize decreases.

FIG. 1B—Outline of Basin during Eocene and early Oligocene time. Geanticlinal ridge emergent throughout. Tectonically high area to east is below sea level. Hollow arrows show direction of fining. Deformation of basin in progress.

sises the unity of the Eastern Basin. The eastward coarsening of the Upper Cretaceous sediments clearly indicates the existence of a sediment-producing area east of New Zealand.

Waipawan Stage

The Waipawan Stage consists of homogeneous, grey, muddy siltstones, red shales, and white to green bentonitic mudstones in Hawke's Bay. The coarsest strata are again to be found in the Raukumara Peninsula. In the Wairarapa, limestones with very thin layers of bentonitic mudstones, greensands and alternating coarse and fine sandstones belong to this stage (Waterhouse and Bradley, 1955). The fine-grained, calcareous deposits may be traced into Marlborough (Macpherson, 1946, 1949; Wellman, 1955), while in North Canterbury the occurrence of glauconite (Wilson, in press) indicates shallowing in the basin.

The tendency in the Upper Cretaceous strata of Hawke's Bay and the Wairarapa to be finer grained in the western parts of those land districts than in the eastern parts is not maintained in the Paleocene. On the contrary the facies changes indicate a landmass in the west and it is very probable that the remainder of the New Zealand Geanticline was emerging at that time. The fine-grained Cretaceous rocks on the crest of this emerging Geanticline provided the source rock for the fine-grained Eocene sedimentation of Hawke's Bay.

Mangaorapan Stage

The Mangaorapan Stage shows very much the same lithologies as the preceding Waipawan except that the sediments are generally finer grained. Green to red bentonitic, silty mudstones occur in the Raukumara Peninsula (M. C. Pick, pers. comm.) while still finer grained, greasy, green grey, bentonitic foraminiferous mudstones prevail in Hawke's Bay (Kingma, 1957a). In the Wairarapa, limestones are to be found together with green to red mudstones, flints, and some alternating fine sandstones and siltstones (Waterhouse and Bradley, 1957; and pers. observ.). The limestones – the Amuri Limestone – can be traced into Marlborough (Macpherson, 1946, 1949; Wellman, 1955), while in North Canterbury glauconitic silty mudstones (Wilson, in press) again point to shallowing.

Heretaungan and Porangan Stages

The Heretaungan deposits are very similar to those of the Mangaorapan, with green and red, bentonitic mudstones in the Raukumara Peninsula, and great thicknesses of greenish grey, bentonitic mudstones very rich in Foraminifera have been observed by the author in Hawke's Bay. Limestones, green to red mudstones, and some small patches with alternating fine-grained sandstone and siltstone are to be found in the Wairarapa, while the Amuri Limestone continues through this stage in Marlborough. In North Canterbury glauconitic siltstones and sandstone, sometimes with bentonitic horizons, form the transition between Marlborough and the carbonaceous and glauconitic, shallow water deposits further south (Wilson, in press).

In the Porangan stage these different types of sedimentation continued, except in Hawke's Bay where a marked fining is to be observed, providing the very greasy, montmorillonite bentonites of that district (Kingma, 1957a). Porangan sands have been found in North Canterbury (Mr D. Hamilton, pers. comm.) merging into shallow-water, carbonaceous sands further south.

The conclusions that may be drawn from the Dannevirke sedimentation are that fine-grained sediments prevailed in the northern part, mudstones in Hawke's Bay, limestones and flints in the central part and arenaceous sediments in the southern part of the basin. The thicknesses of the fine grained sediment is greatest in the Hawke's Bay – Wairarapa areas, and the lack of coarse material probably indicates that the Geanticline was still low during Dannevirke time. It may here be emphasised that again, just as in the Cretaceous, the central part of the New Zealand Geanticline had a low relief when compared with the probably much higher reliefs of the southern and northern parts. The continuity of the Eastern Basin is again well exhibited, this time by the presence of the characteristic Amuri Limestone and flint beds both in the Wairarapa and in Marlborough. The deposition of such a sediment requires a number of very special conditions and its occurrence both in the Wairarapa and Marlborough strongly suggests the continuity of the conditions in the central part of the Eastern Basin.

Bortonian Stage

Glauconitic sandstones, and green to red bentonitic shales, form the sediments of this stage in the Raukumara Peninsula (M. C. Pick, pers. comm.; pers. observ.). Towards Hawke's Bay a distinct facies change takes place (Kingma, pers. observ.), the sediments, becoming much finer grained and marly, consist of light grey mudstones and some fine grained, light coloured, muddy siltstones weathering into bentonitic ooze. Some very pure bentonites occur in the eastern part of that district. These Bortonian marly mudstones differ from the Amuri Limestone only in their smaller calcium carbonate content, but are otherwise comparable. They may be traced as marly mudstones with Amuri Limestone horizons in the Wairarapa and Marlborough where they merge with the Amuri Limestone, until in North Canterbury again glauconitic silt mudstones and greensands predominate (Wilson, in press).

Kaiatan and Runangan Stages

These stages have not been fully traced throughout the basin because of their small thickness and the considerable faulting, they are as a result frequently faulted out of sight, but with more detailed surveys they will probably come to light. The sedimentation follows the same order as the preceding Bortonian stage and consists, apart from mudstones, also of limestones and sandstones with or without glauconite. In Hawke's Bay, white very calcareous marly mudstones occur which may be traced into the Wairarapa where they form conformable sequences from the Bortonian onward; they have not yet satisfactorily been recognised in Marlborough. Part of the Amuri Limestone probably belongs to these stages. Brown grey, marly, glauconitic mudstones prevail in Canterbury (Wilson, in press).

From the distribution of strata carrying index Foraminifera for the Arnold Series the conclusion may be drawn that in the northern as well as in the southern parts of the Eastern Basin definite shallow conditions existed. In the extreme north small, locally deeper basins may have occurred where finer sediments were deposited. In the central part of the basin, sedimentation conditions must have been considerably deeper. The similarity in lithology between the Bortonian of Hawke's Bay, Wairarapa, and Marlborough again shows beyond any doubt the unity of the basin. It also indicates that the central part of New Zealand was too low to produce any coarse sediment.

Whaingaroan Stage

This stage contains impure limestones, sandstones, and calcareous mudstones in the upper part of the Mangatu Formation (Ongley and Macpherson, 1928) in the Raukumara Peninsula (M. C. Pick, pers. comm.; pers. observ.), while in Hawke's Bay a distinctly light coloured mudstone is characteristic (Kingma, 1957a). This mudstone, sometimes with light coloured dense limestones which resemble the Amuri Limestone, continues into the Wairarapa. Its continuation into Marlborough is to be sought in the Weka Pass stone, and possibly in the very top of the Amuri Limestone. In north and central Canterbury, limestones (a local facies of the Amuri limestone known as the Amberly Limestone (Wilson, in press)), and foraminiferal chalks are conspicuous deposits in this stage, while further southward glauconite becomes increasingly prevalent.

The Whaingaroan is the last stage in which fine grained strata are deposited on a large scale in the central part of the Eastern Basin, as in later stages silts and sandstones are the predominant deposits indicating that the relatively quiet conditions of the lower Tertiary had come to an end. Throughout the Eocene and part of the Oligocene the situation has been that the Eastern Basin had rather shallow northern and southern extremes, while fine grained sediments were deposited in its central, apparently deepest part and where, to conform with this sag, the central part of the New Zealand Geanticline had come above sea level last and remained a tectonically low area when compared with the parts further north and south (fig. 1B).

Duntroonian and Waitakian Stages

The sediments of these stages are markedly coarse in the northern part of the Eastern Basin, and certainly coarser than in Hawke's Bay (Ongley and MacPherson, 1928; Lillie, 1953; Kingma, 1957a). In the Raukumara Peninsula sandstones and siltstones sometimes banded and often glauconitic may be traced southward into Hawke's Bay and Dannevirke, where a rather varied shelf facies of medium grained, glauconitic sandstones and glauconitic siltstones is widespread (Lillie, 1953; Kingma, 1957a). Pale grey mudstones and light coloured limestones, together with thick sequences of alternating sandstones and siltstones in the Wairarapa indicate that the Whaingaroan type of sedimentation continued. The sediments may be traced into the limestones and marls of Marlborough and North Canterbury. Southward the limestones prevail, becoming more and more glauconitic (Wilson, in press). From Marlborough northward faulting severely affected the basin and the effects of these movements are reflected in the varied lithology and irregular thicknesses of these stages (fig. 2A).

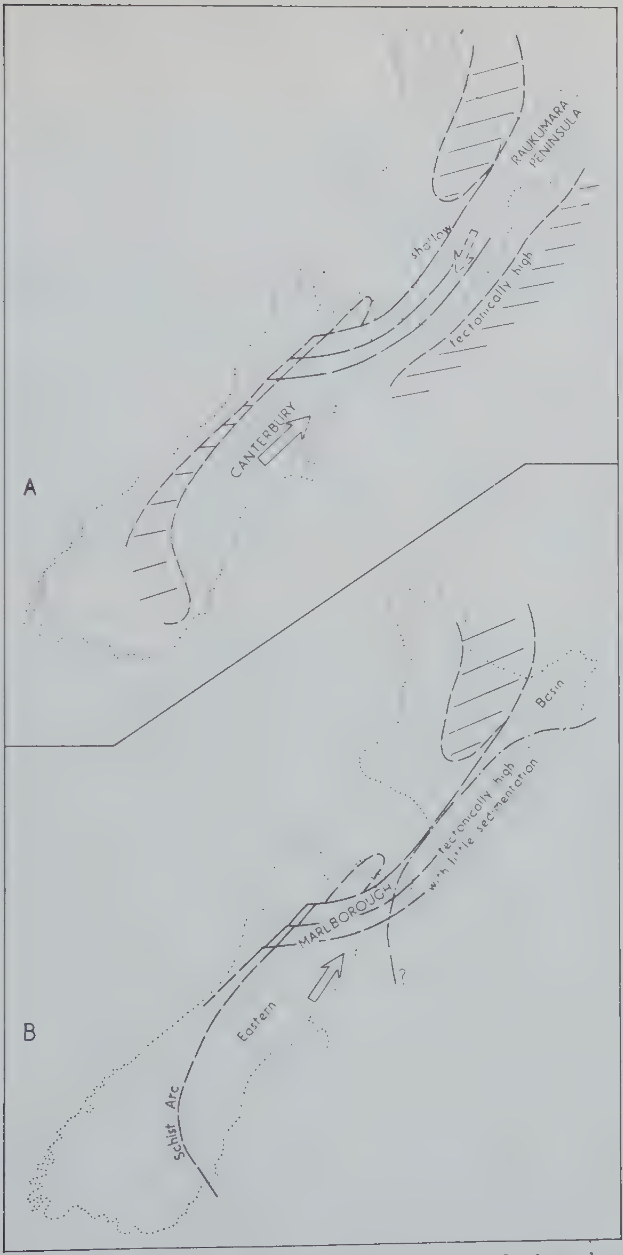

FIG. 2A—Outline of basin in middle Oligocene. Deformation further advanced. Transcurrent fault system developing. Central part of geanticlinal ridge again low.

FIG. 2B—Situation similar to middle Oligocene. Tectonically high area to east of basin has extended further west virtually dividing the basin in a northern and southern part. Strong suggestions for marine connection between southern part of Eastern Basin and western side of geanticlinal ridge through low, central part of New Zealand.

Pareora Stages

Pareora strata have recently been found in the Raukumara Peninsula (M. C. Pick, pers. comm.), consisting of alternating sandstones and siltstones and also mudstones. In Hawke's Bay no sedimentation took place in Pareora time, the area apparently having been high at the time (Kingma, 1957a). In the Wairarapa siltstones with definite Pareora index Foraminifera have now been recognised (G. H. Scott, pers. comm.). In Marlborough and North Canterbury the Pareora stages are represented by marls, limestones and sandstones which are particularly well developed in North Canterbury (Wilson, in press). The sedimentation in these stages appears to be the direct continuation of the Waitakian sedimentation, differing only in the fact that the Hawke's Bay area was brought above sea level. The microfaunal similarity between the sediments of Pareora age in Marlborough and the Wairarapa on the eastern side and the west coasts of New Zealand on the western side, would suggest that the Eastern and Western basins were connected at the time. This would then suggest that, apart from the high area in Hawke's Bay the central part of New Zealand was again low and even below sea level (fig. 2B).

Southland Stages

The Altonian Stage is represented in the Raukumara Peninsula by massive sandstones and by thick sequences of graded bedding as part of the Ihungia Formation (Ongley and Macpherson, 1928). This may be traced into the Dannevirke and Wairarapa areas (Lillie, 1953) while in Marlborough conglomeratic sequences are to be found (Macpherson, 1952; King. 1937). In North Canterbury silty sandstones occur, but further southward no Altonian sediments have been recorded. The Clifdenian and Lillburnian stages follow very similar trends to those observed in the previous Altonian, the lithologies being almost entirely similar.

The Waiauan stage has a slightly larger distribution and occurs further south into Canterbury as the result of a very shallow, marine deposition. Northward in the Eastern Basin the Waiauan is well represented (Kingma, 1957a; Ongley and Macpherson, 1928).

The lithologies in the Southland stages are remarkably similar and can be recognised over large distances; the thicknesses involved, however, are extremely different and provide a feature which makes it so difficult to correlate these stages in the field. The differing thicknesses also indicate the marked tectonic unrest of the time. The large transcurrent fault zones that were already active then virtually tore the Eastern Basin to threads creating depressed areas in one place and tectonic highs in another. The banded Southland sequences are always to be found in these depressions and the banding is obviously a sedimentary response to tectonic movement (fig. 3A).

Taranaki Stages

The Taranaki stages in the Raukumara Peninsula consist of conglomerates and massive sandstones and siltstones, often changing laterally into alternating systems of graded sandstones and siltstones (Ongley and Macpherson, 1929; pers. observ.). This type of lithology may be followed all along the

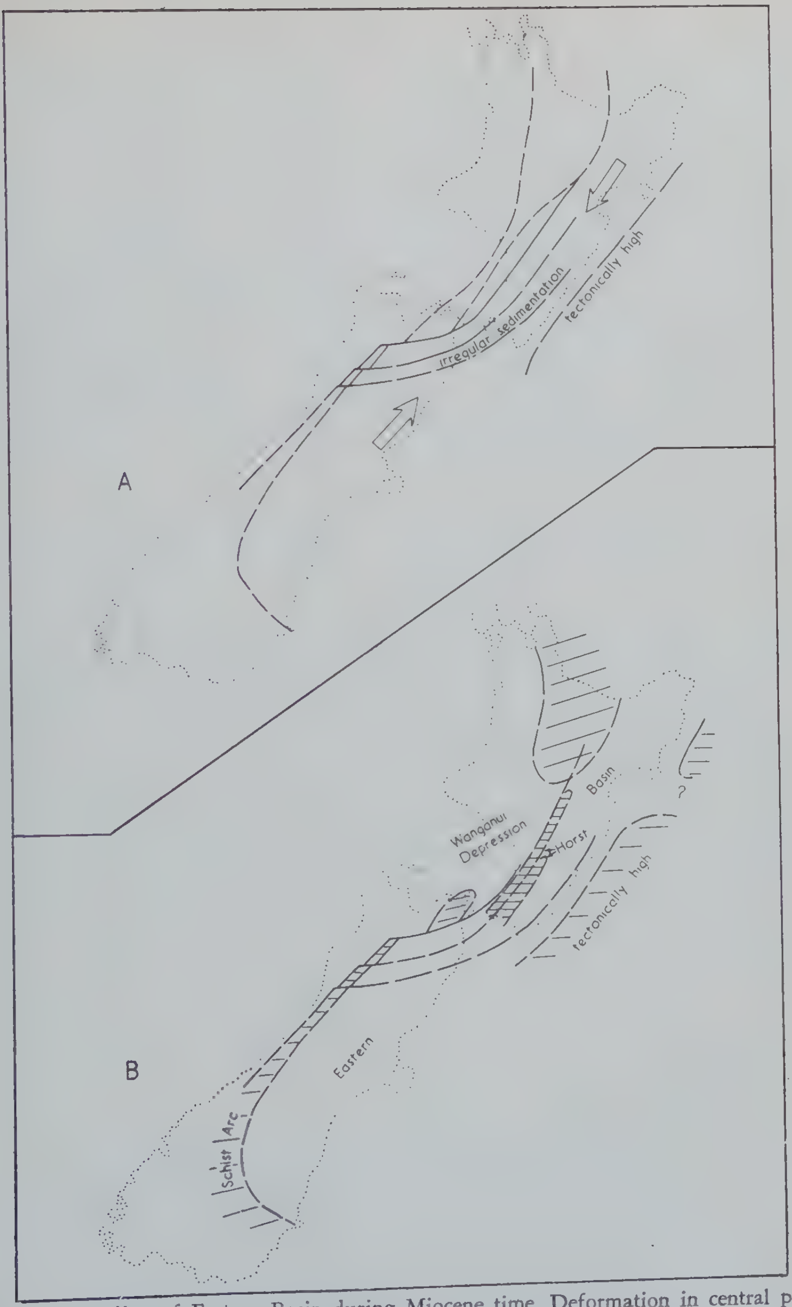

FIG. 3A—Outline of Eastern Basin during Miocene time. Deformation in central part further advanced. Generally irregular sedimentation taking place there. Geanticlinal ridge emergent during most of the Miocene, but submerging again in uppermost Miocene time.

FIG. 3B—During Plio-Pleistocene time submergence continues resulting in Wanganui Depression. Deformation of basin still further advanced, appearance of Ruahine-Rimutaka Horst and Kaikoura Fault blocks and extension of Wanganui depression into Hawke's Bay. Basin completely everted by end of the Pleistocene.

East Coast (Kingma, 1957a), throughout Wairarapa (Lillie, 1953), Marlborough (G. J. Lensen, pers. comm.), and into North Canterbury. The main stage is the Tongaporutuan, which can be divided on micropaleontological grounds into three substages (Kingma, 1958). This subdivision has been found throughout the Eastern Basin and again emphasises the unity of the basin. The sedimentation during the Miocene was thus irregular in thickness and facies. The southern part of the basin was undoubtedly shallow (Suggate and Wilson, 1958), while in the northern part shelf sediments alternated with basin deposits showing graded bedding. The lack of regional conglomerates in the Miocene strata of Hawke's Bay and Wairarapa again indicates that the geanticline was not far above sea level, and may have been below sea level during part of that time as it has been found that the Southland microfaunal assemblages of the Raukumara Peninsula are similar to those found in the Southland stages of the west coast of the South Island (N. de B. Hornibrook, pers. comm.).

Wanganui Stages

The strata of the Wanganui stages may be divided into terrestrial and marine deposits, the later being found mainly in the central part of the Eastern Basin (Lillie, 1953; Kingma, 1957a), and on the submerged geanticline in the Wanganui Depression (Fleming, 1953), (fig. 3B). The tectonic upheaval was considerable at the time resulting in the further disruption of the Eastern Basin and the appearance of an elongated greywacke ridge, the Ruahine-Rimutaka Horst (Kingma, 1957, a, b). This horst acted as a barrier, very similar to the earlier New Zealand Geanticline, between the Eastern Basin and the sedimentation system in the western part of New Zealand. The geanticline itself was depressed, showing again the general trend that the central part of New Zealand was tectonically low. The Eastern Basin was further torn apart in the Hawke's Bay area and substantial sedimentation took place in the ruptured area and along the Ruahine-Rimutaka Horst.

CONCLUSIONS ARISING FROM THE STUDY OF THE SEDIMENTATION OF THE EASTERN BASIN

Four important divisions may clearly be recognised in the sedimentary history of the Eastern Basin:

1. An Upper Cretaceous - earliest Tertiary part, when the Eastern Basin was formed as a separate basin of deposition, except in the central part where geosynclinal conditions from the earlier New Zealand Geosyncline still persisted. The sediments in the Eastern Basin were rather coarse where the basin flanked the New Zealand Geanticline, i.e. in the northern and southern parts, while the grain size decreased towards the central part.

2. An Eocene - earliest Oligocene part, where the Eastern Basin was bordered on the west by a fully emerged New Zealand Geanticline, but which was still low in altitude in its central part. The characteristic deposits in this basin are fine-grained, calcareous, bentonitic mudstones in the central part of the basin, coarsening north and southward.

3. An Oligocene-Miocene part when the sediments generally coarsened. Local facies changes indicate that faulting was well under way and displacement of the segments in between transcurrent fault zones was in progress. Judging from facies changes the central part of the New Zealand Geanticline was again low.

4. A Plio-Pleistocene part when the basin generally broke up leaving irregular depressions into which coarse and fine grained strata from the adjoining highs were laid down.

The sediments were fine grained in the central part of the Eastern Basin in the uppermost Cretaceous and Eocene, while the relief of the New Zealand Geanticline was low at the time. This particular feature persisted throughout the Cenozoic and may be seen today in the Wanganui Depression. In post-Oligocene time the central part of the Eastern Basin was tectonically too disturbed to have fine grained sediments, but persistence of characteristic microfaunas and persisting tectonic features from Wairarapa into Marlborough emphasise the unity of the basin. The fine grain size of sediments is again shown by the lack of conglomerates in the Oligo-Miocene sequence of the Hawke's Bay and Wairarapa areas.

Structural Implications

In Upper Cretaceous and Lower Tertiary times the Eastern Basin seems to have been deeper in the central part than in the extremities further north and south. This lends support to the author's hypothesis (Kingma, 1959a, b) that the New Zealand Schist Arc was being wrenched off the northern portion of the New Zealand Geanticline. In such a process a considerable amount of stretching must have taken place at and around the area where the wrench-off occurred. The generally deeper conditions in the central part of the Eastern Basin and the generally lower relief of the New Zealand Geanticline in the area of the present Wanganui Basin are probably best explained as the result of such stretching, and consequent sinking movements. In the early stages of this sinking sedimentation in the central part of the Eastern Basin was generally quiet, but as the Alpine Schist Arc progressed, the central part became more and more deformed. The deformation along fault zones was accompanied by the development of basins and horsts.

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COASTAL SURFACE CURRENTS AROUND NEW ZEALAND

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Summary

Recoveries of drift cards from releases made off the New Zealand coast in the twelve months preceding July 1954 demonstrate the existence of a coastal circulation, dependent on the surrounding oceanic currents and the topography for its basic pattern.

INTRODUCTION

Outlines of surface ocean current movements in the southwest Pacific Ocean and the Tasman Sea around New Zealand have been given by a number of authors (Deacon, 1937; Barlow, 1938; Fleming, 1944; 1950; and Dell, 1952). To obtain data on the detail of these movements in the near vicinity of New Zealand and to obtain some additional information on oceanic movements particularly in the Tasman Sea, plastic-enclosed drift cards (Olson, 1951) were employed. Some 12,000 cards each 3 in. × 5 in. were dropped in groups of ten at release points within an area bounded by the east Australian coast and longitude 145°E in the west, by longitude 170°W in the east and extending from latitude 20°S to 55°S. Releases were made from merchant, naval, and government vessels at points on the following pattern: each 5 miles to 40 miles clear of the New Zealand coastline; then each 40 miles between 40 and 200 miles offshore; then each 200 miles to the limit of the area. On Tasman Sea routes a similar concentration of release points was used on approaching or leaving the Australian coastline. Additional releases were made from vessels on coastal passages around New Zealand.

Four quarterly groups of releases were planned, the first centering on 31 July 1953. Ship sailings, on which releases depended, were spread from mid-June to mid-September; a similar spread in time had necessarily to be accepted in successive quarters to secure sufficient geographic coverage.

Recoveries of cards were entirely made through the cooperation of private individuals who were asked to fill in details of time, place, and circumstances of finding of cards returned.

DATA ON DRIFT CARD RECOVERIES

Cards were returned from numerous points on the New Zealand and Australian coastlines and the apparent remoteness of some localities did not prevent recoveries from such areas being made. Cards were returned in relatively large numbers until 1956: since this year, less than twenty cards have been received.

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Cards from both the Australian and New Zealand coastal drops have been recovered in moderate numbers (table 1). Only one card was returned from any oceanic release point more than 160 miles offshore and few from points more than 80 miles offshore (Fig. 1).

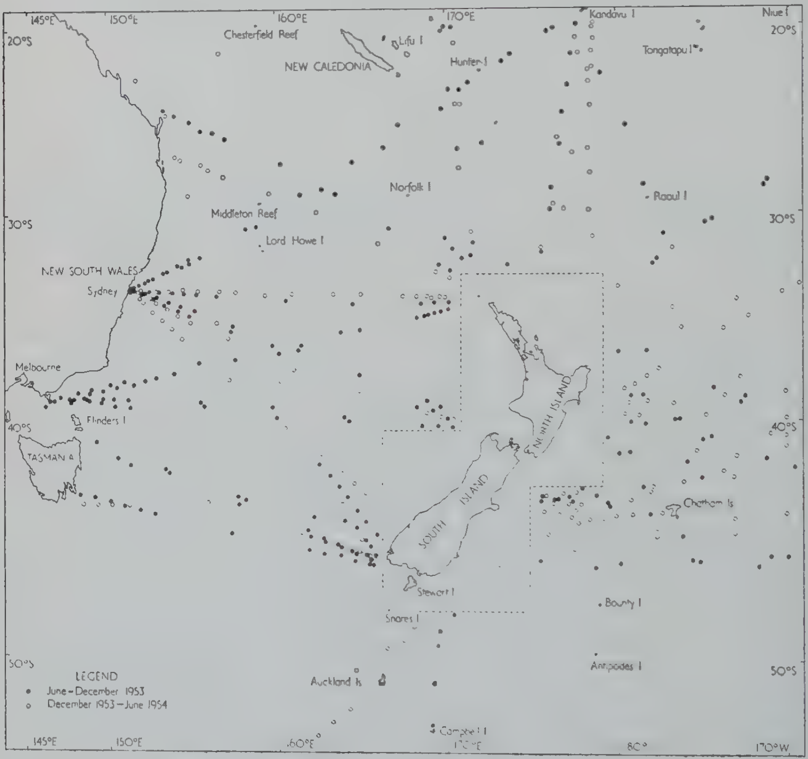
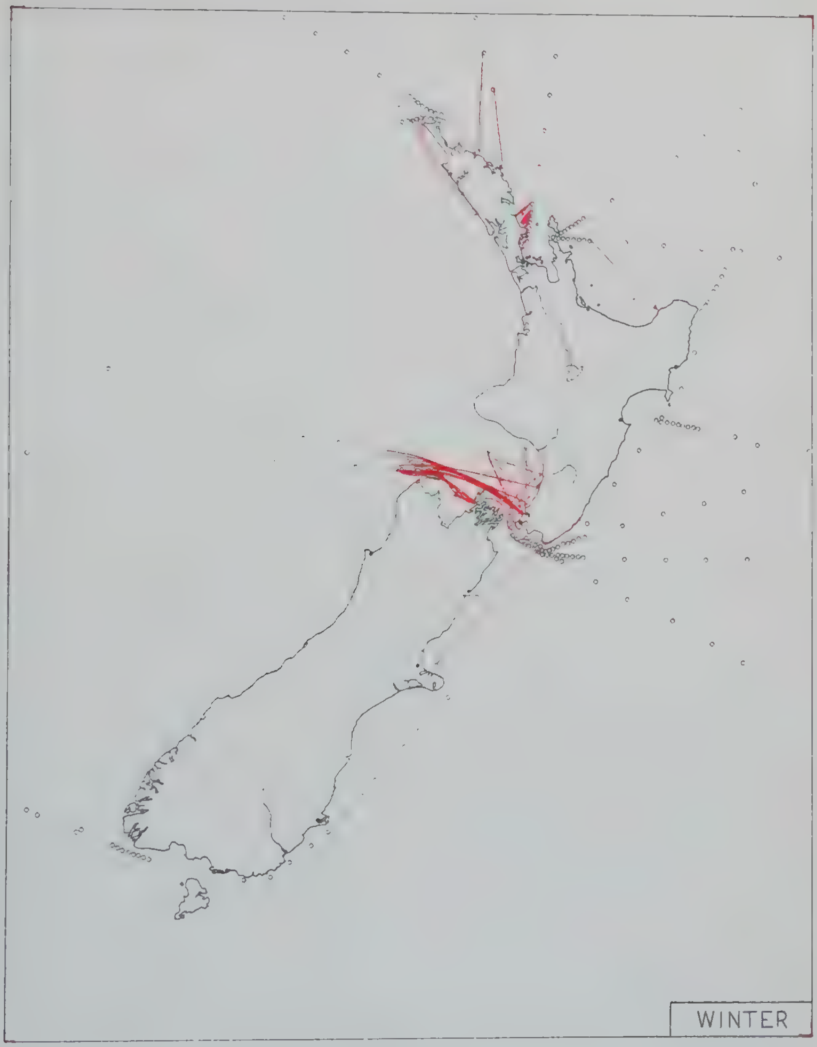


FIG. 1—Oceanic Release points of drift cards.

TABLE 1—Recoveries of Drift Cards

Area of Release		Number Released	Number Recovered	Per Cent Recovered
New Zealand, up to 160 miles offshore	......	6,820	357	5·2
Australia, up to 160 miles offshore	......	2,310	53	2·3
Oceanic, more than 160 miles offshore	......	2,930	1	......

The recoveries of cards are shown in figs. 2, 3, for releases made in each quarter.

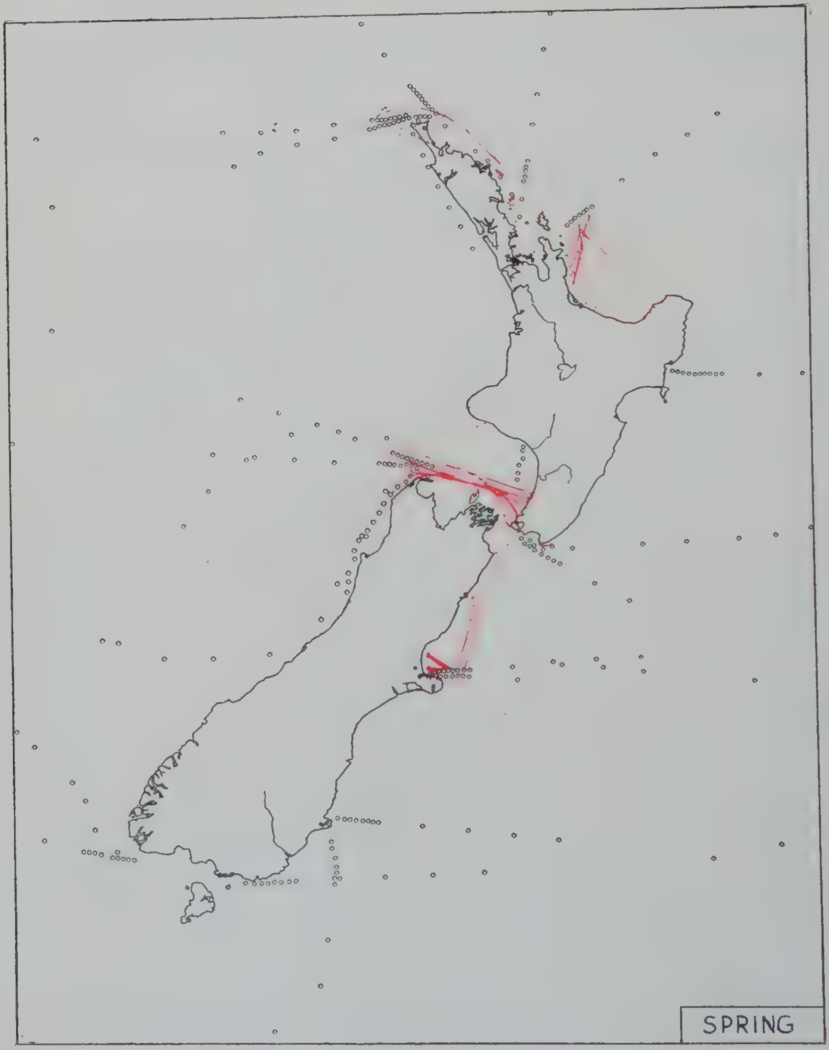


(Fig. 2a)

FIG. 2—Tracks of drift cards recovered on the New Zealand coast; (a) from Winter 1953 releases (mid-June – mid-September); (b) from Spring 1953 releases (mid-September – mid-December). Approximate shortest tracks between release and recovery points are shown.

1. *Recoveries from Oceanic Releases*

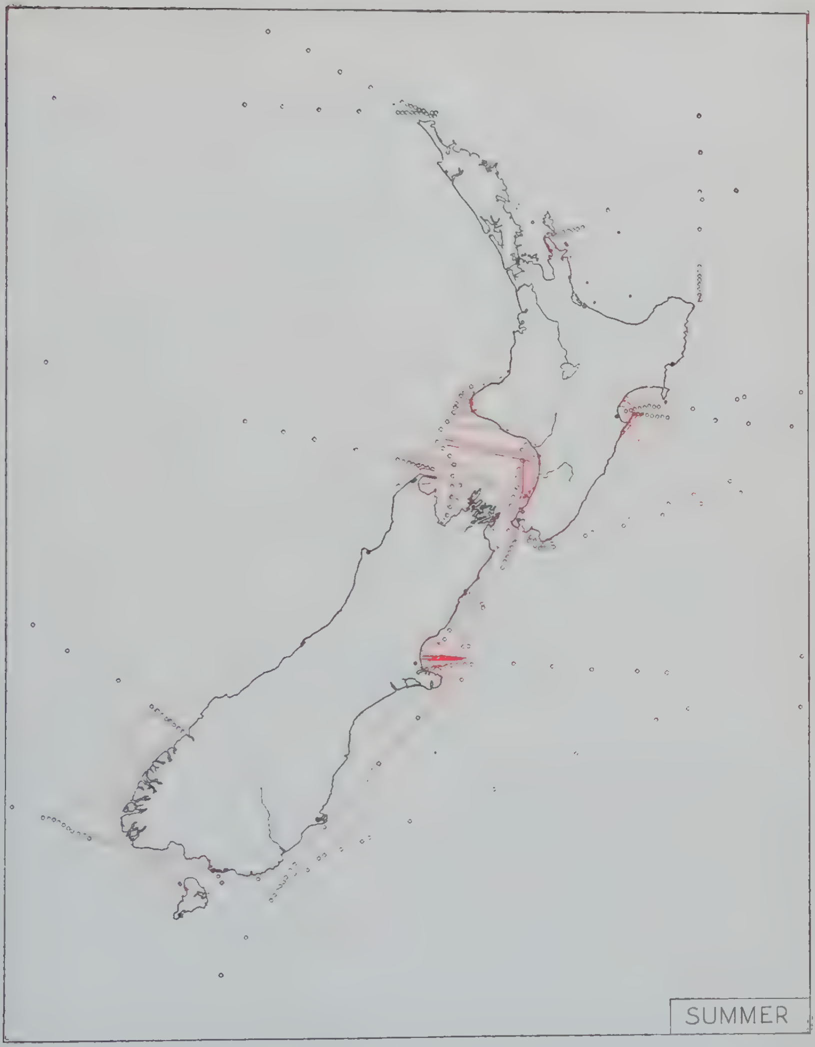
The one recovery made from release points more than 160 miles offshore, was a card dropped 170 miles east of Otago Harbour which was found at Te Whanga Lagoon in the Chatham Islands 400 miles further east.



(Fig. 2b)

2. *Australian Coast Recoveries*

The Australian coast releases were principally made approaching or leaving Sydney. Recoveries were made at points up to 50 miles north and 100 miles south of Sydney but many of the 53 cards recovered were from the near vicinity of the release points. Further analysis of the results has not been made.

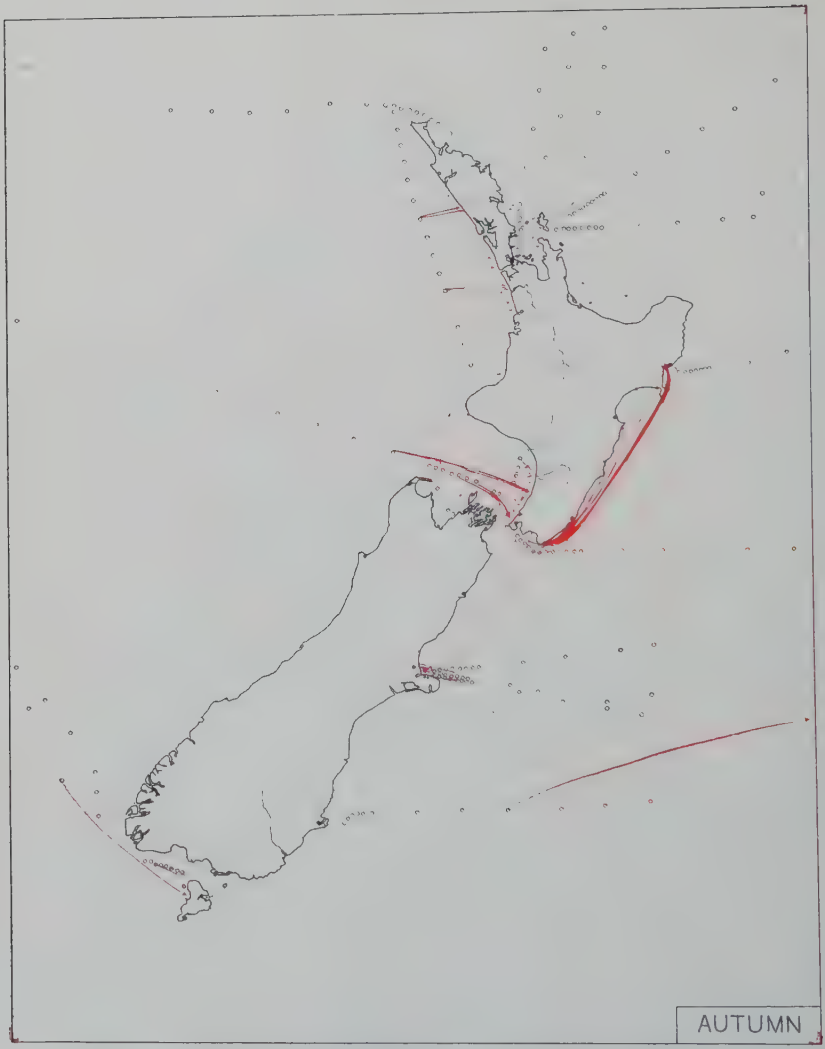


(Fig. 3a)

FIG. 3—Tracks of drift cards recovered on the New Zealand coast: (a) from Summer 1953–4 releases (mid-December–mid-March); (b) from Autumn 1954 releases (mid-March – mid June). Approximate shortest tracks between release and recovery points are shown.

3. *New Zealand Coast Recoveries*

The pattern of recovery of cards dropped in New Zealand coastal waters did not vary markedly from quarter to quarter. When the release-to-recovery tracks are plotted for the whole year, two major points emerge.



(Fig. 3b)

Firstly that recoveries have been made mainly from releases in a narrow coastal belt, extending approximately 100 miles offshore from the west coast and 50 miles offshore from the east coast. Secondly that the pattern of drift revealed is consistent in that with the exception of one area the drifts recorded off any one portion of the coastline are all in the one direction. The exception occurs in the Cape Farewell to Manukau Harbour area.

The drift tracks indicate surface water movements as follows (localities are shown in fig. 4):

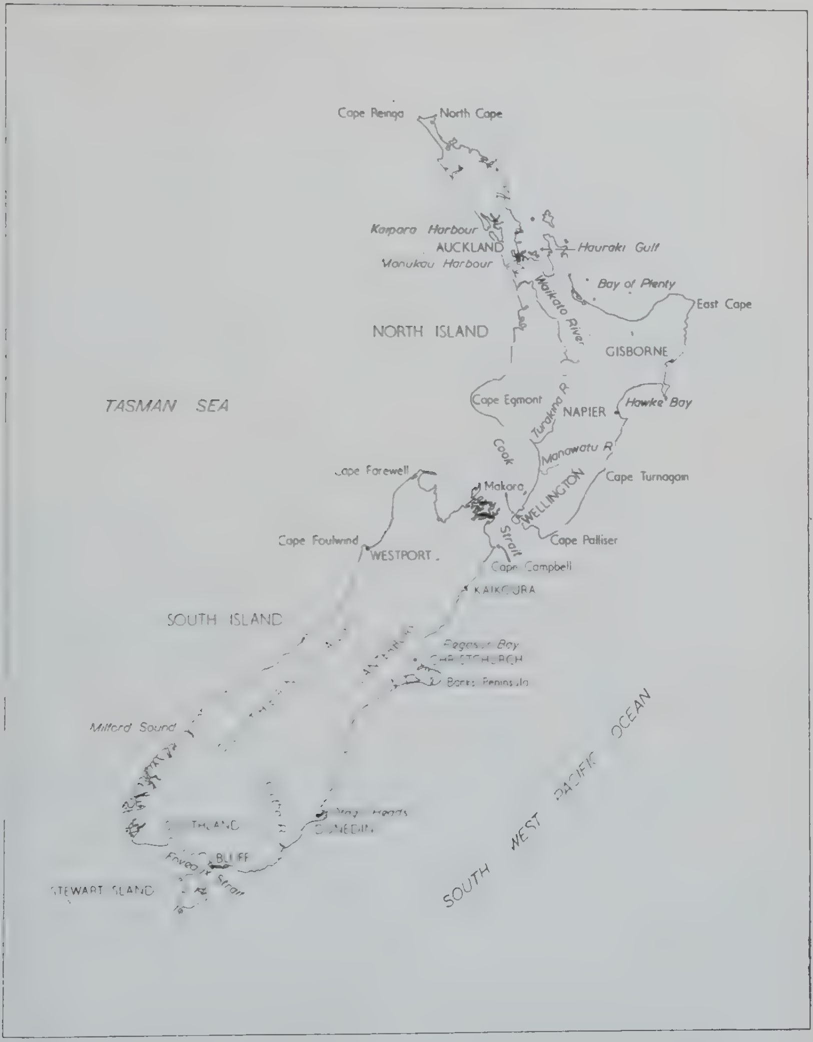


FIG. 4—Localities mentioned in text.

1. *East Auckland Current:* From North Cape down the Auckland east coast to the Bay of Plenty.
2. *West Auckland Current:* From north and west of Cape Reinga down the Auckland west coast towards Kaipara.
3. *Southland Current:* From the West Coast Sounds region and to the west, east through Foveaux Strait and up the east coast of the South Island.

4. *Canterbury Current:* From Banks Peninsula north past Cook Strait and the east coast of the North Island as far as Gisborne.
5. *Westland Current:* From west coast South Island up to Cape Farewell and less markedly north along the Cape Egmont coastline and beyond towards Manukau Harbour.
6. *D'Urville Current:* From off the western entrance to Cook Strait, east into the Strait and thence south towards the narrows of Cook Strait. The drift sometimes continues less markedly to the east along the South Wellington coastline.

The West Auckland, Southland, Canterbury, and Westland Currents have been described and named by D. M. Garner (1960), from an analysis of the coastal hydrology for the year 1955. The East Auckland Current and the D'Urville Current are defined and named here: these movements were also discussed by Garner (loc. cit.).

EARLIER DRIFT INVESTIGATIONS

From 1920 to 1922 fortnightly releases of ballasted drift bottles were made off Otago Heads (Thomson, 1922). From 438 releases, 51 recoveries were made. Except for local strandings near the release points all bottles drifted to the north and north east: 34 stranded on the east coast between the release point and south of Banks Peninsula; 7 on the east coast north of Banks Peninsula; 1 at Lyall Bay, Wellington; and 3 at the Chatham Islands, 600 miles N.E. of the release point. Bottles released less than 5 miles offshore tended to strand near the release point.

Manuscript logs and notebooks held at Portobello Marine Biological Station show that additional recoveries were made later, 2 more bottles being found at the Chatham Islands, and 1 at Cape Turnagain on the North Island east coast. A further 46 bottles were released off Otago Heads in 1922 and 132 in 1928, all those recovered being found locally (Marine Dep. Annual Report, 1929). In the winter of 1926, 213 bottles were dropped at various points in Foveaux Strait and 48 more early in 1927. Fifteen were recovered from the nearby coasts of Southland and various islands in the strait. Six drifted north to points on the South Island east coast from the Clutha River to Banks Peninsula: one was recovered from the Chatham Islands.

In 1918 and 1919, the New Zealand Marine Department released 25 wooden floats, each a ballasted frame 3 ft square, in an area centred 30 miles west of Cape Farewell. The object was to simulate the drift of mines breaking loose from their moorings in a minefield laid in this area. Twenty of the floats were recovered between Makara and Turakina on the Wellington west coast, 7 between Cape Egmont and Manukau Harbour and 2 on Farewell Spit. From 1951 to 1952, 106 bottles were dropped from HMNZS *Lachlan* and 12 bottles were recovered. These bottle and float drifts are shown in fig. 5.

Dell (1952) analysed the recoveries on the New Zealand coastline of drift bottles released by H. C. Russell (1894). Bottles released in the Southern Ocean west or south of Australia were recovered from Cape Reinga - Cape

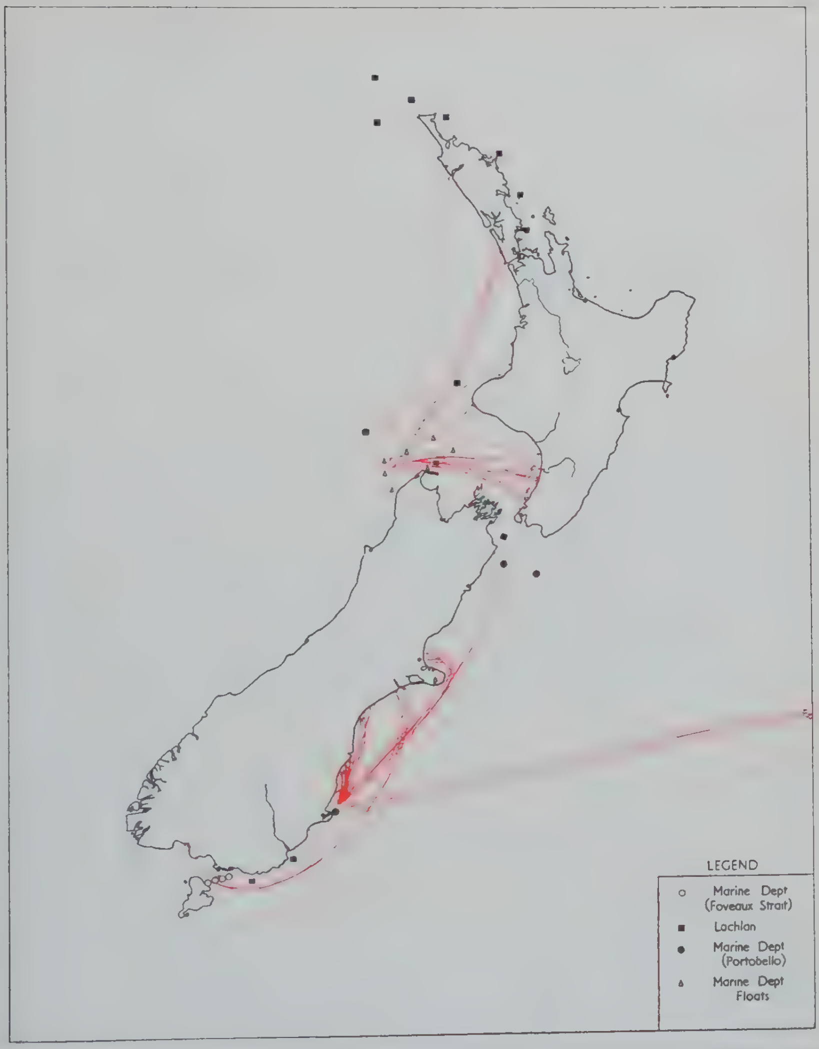


FIG. 5—Tracks of drift bottles and floats released in New Zealand coastal waters 1918–1952. Approximate shortest tracks betwen release and recovery are shown.

Egmont, 7; Cape Egmont - Wellington, 7; Cape Farewell - Cape Foulwind, 2; and on eastern coasts, North Cape - Auckland, 1; Stewart Is. - Banks Peninsula, 5; Chatham Is., 1. Bottles released in the Tasman Sea off the New South Wales coast and well west of New Zealand were recovered from Cape Reinga - Cape Egmont, 20; Cape Egmont - Wellington, 7; Cape Farewell - Cape Foulwind, 2; and on the east coast, North Cape - Auckland, 3; Auckland - East Cape, 2; Cape Campbell - Banks Peninsula, 2; Stewart Is. - Banks Peninsula, 3.

Dell's interpretation of these movements is as a divergence off the southwest coast of New Zealand of West Wind Drift mixing with a counter clockwise Tasman Sea circulation into two streams, the major movement up the west coast of both islands, the lesser along the east coast of the South Island.

Recoveries from Russell's oceanic releases present some points of difference from the present series. Most notable is the concentration of recoveries in the Cape Egmont - Cape Reinga region at the end of drifts which must have had strong northerly and easterly components. This emphasis on possible northward movements along the west coast of the North Island towards Cape Reinga is absent from the present results and in part negatived by the southward movement from Cape Reinga.

DISCUSSION

The Weather Pattern

Between the southeast trade zone north of 30°S and the strong westerly flow in the "roaring forties" south of 45°S lies an intensive high pressure belt. This belt which covers most of New Zealand except the southern portion of the South Island, consists in detail, of a series of semi-permanent anticyclones which move slowly to the E.N.E. crossing the country at the rate of roughly one per week. As each moves off New Zealand to the east, northwesterlies develop and a low pressure trough follows frequently with a cold front. The front is followed by a wind change from northwest to west, southwest, south, or southeast. Watts (1947) to whom this analysis is due, tabulates the major controlling influences (table 2) on the climate of three regions.

TABLE 2—Major Controlling Influences on the Climate of the Three Regions of New Zealand (from Watts 1947)

Controlling Factor	Northern Region	Central Region	Southern Region
	%	%	%
Northwesterly, westerly, and southwesterly	39	41	37
Southerly to easterly	14	21	21
Anticyclone	28	19	20
Moist northerly	5	4	3
Depressions, warm fronts, stationary front	14	15	19

Westerlies and northwesterlies converge in the northern Cook Strait area into northwesterlies and northerlies, later these recurve and extend over Canterbury as northeasterlies.

The local variations in number of gales per year, wind miles run per day, and most frequent direction of wind at 9 a.m. are shown in fig. 7. Watts comments as follows: "Thus it will be seen that the surface configuration of the country and the orientation of the Straits play an important part in the formation of our weather. [Air] Currents from the northwest, south, or southwest are all concentrated and controlled by the shape of central New Zealand [the Cook Strait Area]. Further south, northwesterlies, westerlies, and southwesterlies are diverted by the Alps into a strong [air] flow through Foveaux Strait and a similar number of gales per year is general through both Cook and Foveaux Straits. Forty per cent of the year under westerly control does indeed have the most outstanding influence on average conditions: but the despised twenty per cent of southerly to easterly control is the determining factor of several phenomena."

Significance of Drifts

The movement of drift cards will be influenced by several factors: (a) the regional oceanic surface currents and the effects of the disposition of land-masses on them; (b) the circulation induced by local wind systems; (c) the effect of the wind on the few centimetres depth of water in which the card is floating; and (d) the immediate effect of the wind on the card.

The direct effect of winds on the cards themselves is probably small. Where cards were observed after dropping, they were seen in some cases to sail before the wind until the surface of the cards became wet when they lay flat in the immediate surface layer of water. It is doubtless true also that in rough water, portion of the card intermittently projects above the water surface. However the forward movement acquired by the card in this way is probably similar to the forward transport of water in the surface few centimetres by the breaking of wave crests and the blowing of spray by the wind. According to Hughes (1956, 1957) recoveries of drift cards from the Eastern Atlantic indicate that the top few centimetres of water travel approximately in the direction of the gradient wind and at 2·2% of its velocity.

The stranding of cards can be a result of several processes. Broadly, a local accident of tide, onshore wind or onshore surface set determines the strandings of cards transported to the near vicinity (see also Garner, in press). Of the cards not recovered a proportion must be similarly transported to the near vicinity but must avoid the accident of stranding, and the balance, excepting sinkings and possible losses due to seabirds, are not brought into onshore situations.

Effect of Local Winds

The release of cards on 10 April 1954 in the Cape Palliser area was followed by a number of recoveries along the North Island east coast as far north as Gisborne during the following 6–8 weeks. Seven cards travelled more than 200 miles north-north-eastward. The weather situation in the release area has been examined for the period immediately following the release date. On the day of release, winds from a northwesterly direction prevailed (10 knots); the following day from the south (10 knots); then on succeeding days; north-east (20 knots); light winds from south and southwest; southeast, southwest and west (15 knots); west and northwest (25 knots); northwest (light); northwest (25 knots); south, southeast; then northeast and northwest (light) for this and two succeeding days; south (15 knots) on 22 April. Similar variation in wind direction occurs for the balance of the month.

The northward component of wind run for the twelve days detailed is approximately the same as the southward component and after this time cards could well have been transported out of the area. Predominant onshore (south to northeast) components are not evident. However, moderate winds (10 knots) from south to southeast are recorded for the area offshore from Hawke Bay and Poverty Bay for the four days preceding the period (2–9 May 1954) in which five cards stranded near Gisborne. Following a similar release of cards off Cape Palliser on 3 October 1953, winds with an onshore easterly component predominated for the rest of the month. No recoveries were made from this release except three cards released near shore which travelled only 16 miles before stranding.

Off the North Canterbury coast the most frequent wind is from the north; no drifts to the south have been recorded in this area.

The net effects of local winds on the surface circulation thus appear to be much less significant than the control exerted by oceanic circulation, except where the added effects of topography are felt as in Cook and Foveaux Straits, or when severe or prolonged gales occur.

Rates of Drift: Rates of drift deduced from the interval between release and recovery will always be minimum rates, for the tracks could well be longer than the straight line or shortest course, and stranding could occur long before recovery.

The fastest drift recorded was 10·2 miles per day (interval to recovery, 22 days); the four next fastest averaged 9 (average interval, 24 days); the next eight averaged 7, and the next thirty cards averaged 4 miles per day (average interval, 60 days).

Most of the drifts reported here took place over long periods, a month being a typical interval between release and recovery. While the uncertainty due to unknown delay in recovery after stranding cannot be avoided the probability is high that the period afloat is very much greater than the duration of any one period of persistent winds from one direction. Thus there is no basis for computing rates of drift in terms of gradient wind velocities.

Inshore Currents: Cards dropped close to shore have naturally exhibited the most obvious response to tidal movements. From the one release point strandings in the immediate vicinity both up and down the coast have occurred from release points up to ten miles offshore.

In Pegasus Bay, card recoveries showed a noticeable difference between the drift from onshore and offshore releases. Cards dropped at points up to 20 miles offshore in the southern portion of the bay have principally been recovered on the Pegasus Bay beaches after a nett drift due west. Cards dropped 20–60 miles offshore have moved northeast parallel to the coast line (as the Canterbury current).

An analysis of 90 recoveries of Thomson's drift bottles released off Otago Heads shows (table 3) that of the bottles dropped further out than 10 miles none was stranded locally and all three recovered had travelled more than 100 miles (two to the Chatham Is.). Conversely recoveries from drops up to 5 miles offshore were all made within 50 miles of the release point.

TABLE 3—Comparison of Distances Travelled by Drift Bottles and Distances Offshore of Release Points off Otago Heads

Distance Offshore of Release Point (Miles)	Distance Travelled (North-east) 0–50 m (%)	50–100 m (%)	100 m (%)
0–5	42	0	0
5–10	35	0	20
10–15	0	0	3

THE COASTAL CURRENTS

Offshore from New Zealand, three principal surface water movements give rise to the coastal circulation. To the south of the subtropical convergence lies the West Wind Drift; to the north of the North Island, the Trade Wind Drift; and west of both islands, the Tasman current broadly north- and east-flowing in the eastern Tasman Sea. The coastal current pattern can be derived from the direct effect of these regional influences, combined with the varying effects of extremes of local winds.

In considering the coastal current pattern generalised in fig. 6 it must be kept in mind that the extent of each of the movements shown is variable. The significance of the pattern is the lack of conflicting directions of movement indicated, but at any one time a particular surface current may be strongly or weakly developed.

The individual components of the coastal current system can be considered separately: *The East Auckland Current*. This movement is strongly indicated by relatively numerous recoveries and arises through the interposition of the north-east coastline across a general westerly and locally south-westerly flow of subtropical water from the Trade Wind Drift, which similarly gives rise to the less strongly developed *West Auckland Current* (Garner, in press).

The *Westland Current* arises as the coastal portion of the modified subtropical water of the counter-clockwise circulation in the Tasman Sea (the Tasman Current). The northern limit of this current is variable, for although the inshore zone tends to round Farewell Spit into Cook Strait yet patently at times the waters off the western entrance to Cook Strait are moving northward to influence the North Island west coast north of Cape Egmont.

Changes in the relative extent of the north-going Westland and the south-going West Auckland drifts along the North Island west coast (apart from any seasonal effect which might occur) probably reflect changes in the local weather situations, strong southerly winds minimising the indraught into Cook Strait and extending the Westland current to the north.

The marked indraught into Cook Strait here named the *D'Urville Current* has been long known to navigators, not least to Dumont D'Urville himself who in 1827 unexpectedly rounded Cape Farewell from the west overnight. "14th January (1827) – At three o'clock in the morning I steered in the direction in which I assumed that we should find Cape Farewell, but at daybreak I realised that during the night the set of the current had carried us in surprising manner to ENE, and we were already a fair distance inside the Strait." (Wright, 1953). This current parallels the direction into which southwesterly, westerly, northwesterly, and northerly winds are funnelled through the Strait by the high land of both islands; westerly and then northerly winds predominate as the Narrows are approached from the west. Under strong and persistent southerly gales, some surface movements to the north have been recorded. However, the very large number of drifts which have been recorded substantiating the surface flow to the east into the Strait and then south to the Narrows marks this current as a near-permanent feature of the Strait. Watts (1947) records 41% of days in the central region of New Zealand on which northwesterly, westerly, and south-westerly winds have prevailed.

In similar fashion, topographic deflection and concentration of winds around the main mountain mass of the South Island through Foveaux Strait, coincides with the topographically restricted channel available for oceanic surface water movements and combines to produce a marked surface flow, the *Southland Current* from the lower west coast of the South Island and the western approaches to Foveaux Strait through the strait and along the south-east coast of the South Island towards and north of Otago Peninsula. This is a branch of the Tasman current with a varying proportion of direct effect of the West Wind Drift. Within this current many drifts north from Otago Peninsula terminated on Pegasus Bay beaches, or south of Banks Peninsula. From the vicinity of Banks Peninsula there is a northward movement of cool sub-Antarctic water, the *Canterbury Current* (Garner, in press). Cards dropped outside Pegasus Bay have travelled north in this current, and drifts show that the movement may extend as far north as Gisborne on the North Island east coast. It is notable that the effects of the northeasterly winds common over north Canterbury do not appear to influence the drifts in the Canterbury Current. All recorded drifts are to the north, none to the south. The boundary between this and the Southland Current is not clearly defined in the present survey. Towards Banks Peninsula the Canterbury Current occupies a more offshore position, than evidenced here for the Southland Current. This northflowing current is bounded to the east by the subtropical southflowing *East Cape Current* (Fleming, 1950; Garner, 1953). This is a movement which lies offshore except where it meets the east coast for some distance south from East Cape, and no drifts have recorded this current, except two from a release 10 miles north of East Cape, the cards from which travelled to points 10 miles and 30 miles south of the Cape. A total of 1,000 cards was released in the East Cape Current.

The drifts of six bottles from offshore positions near the South Island east coast to the Chatham Islands and the similar drift of one card from a position 170 miles east of Otago Heads are confined by the eastward movement in the East Cape Current (Garner, 1957) to the area south of Banks Peninsula. A "Chathams Current" particularising part of the West Wind Drift in this area cannot be erected on the available evidence, which demonstrates the general drift towards the east, which is taking place south of the sub-tropical Convergence.

For the coastal current system as a whole the influence of the oceanic currents is sufficiently strong to maintain the direction of flow in the coastal zone, in the face of opposing winds.

ZONE OF COASTAL CIRCULATION

The limits of the area from which drift cards have been recovered are shown in fig. 6. Cards dropped more than 100 miles off the west coasts or more than 50 miles off the east coasts have apparently little chance of being cast ashore: offshore from this zone the regional influences predominate. This is therefore the zone from which similar drifting material might be expected to strand due to the increasing significance of topographic, tidal and local wind influences. The probability of strandings of drifting material from outside this zone is low.

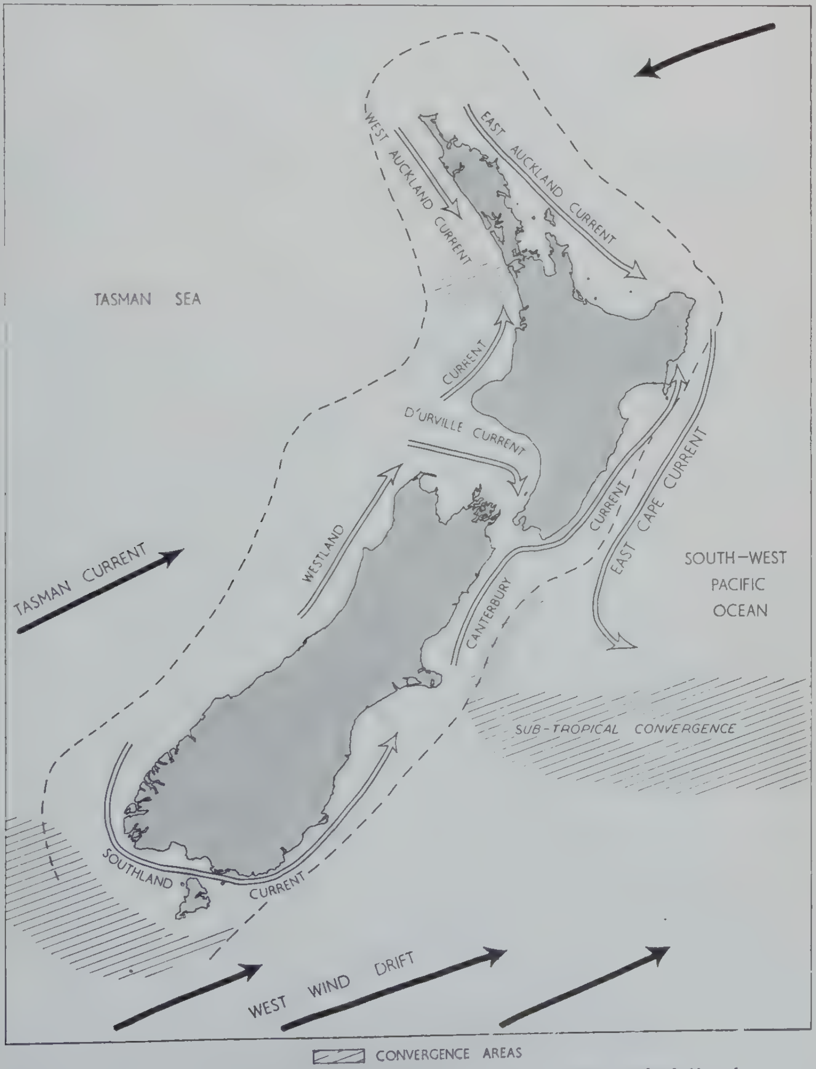


FIG. 6—**Coastal current pattern derived from drift card tracks. Pecked line is seaward limit of the zone from which drift cards have been recovered.**

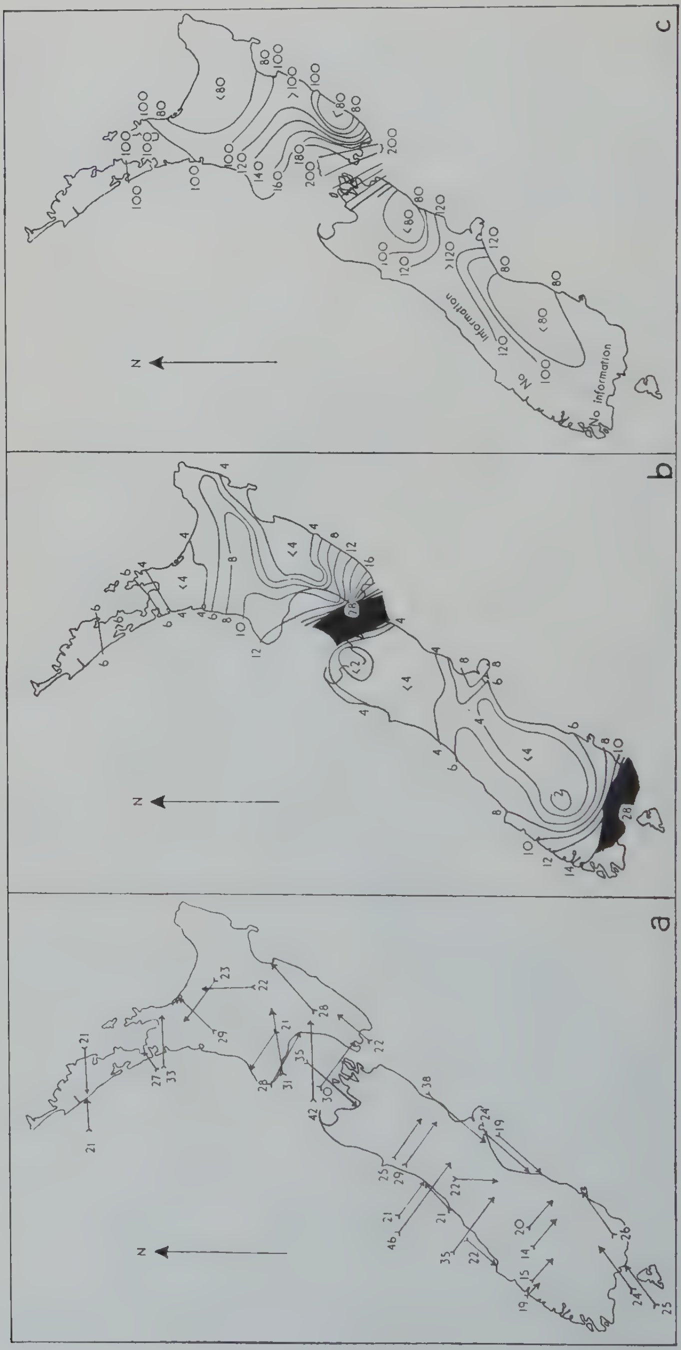


FIG. 7—Meteorological data (reproduced from Watts, 1947): (a) Most frequent wind direction at 9 a.m. (shown as percentage); (b) Number of gales per year; (c) Wind miles run per day.

CONCLUSIONS

The coastal current system around New Zealand driven by three major oceanic surface current systems – the Tasman Current, West Wind Drift and Trade Wind Drift – is conditioned by the orientation of the coastline and modified in places by extremes of local winds. The directions of flow within the system are consistent.

Coastal influence including that of wind and topography is apparently limited to a zone extending round the shore 100 miles wide off the west coast and 50 miles wide off the east coast. Drift cards dropped outside this zone have not been brought on shore and recovered.

The limitations of the method by which the present results are derived should be borne in mind. Longshore drifts from which cards are stranded by a variety of accidents, and longshore drifts with an onshore component are recorded: drifts parallel to the coast, the cards in which are not stranded, similar longshore drifts with an offshore component, and drifts directed offshore are not recorded. These may well be substantial portions of the local current system.

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SURFACE WATER MOVEMENTS IN HAWKE BAY, NEW ZEALAND

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(Received for publication, 26 November 1959)

Summary

A pattern of movement of surface water in Hawke Bay has been indicated by the use of plastic covered drift cards. The results, generalised over a period of three summer months, indicate a strong central flow of water into the bay diverging to move north and south parallel with the shore.

Introduction

The pattern of surface currents in New Zealand coastal waters has recently been described in general terms by Brodie (1960). A more detailed surface drift experiment has enabled a closer examination to be made of the surface circulation in a large, semi-circular bay. Seven hundred plastic covered cards (Olson, 1951) were released from an aircraft over Hawke Bay on 13 December 1958.

Hawke Bay (fig. 1), situated on the east coast of the North Island, New Zealand, is a large indentation extending in a broad crescent from the inner coast of Mahia Peninsula in the north-east to Cape Kidnappers, about 50 miles to the west-south-west. The bay is entirely open to the south-east; the bottom contours are regular and generally parallel the shoreline of the bay. The depth at the entrance is 60 fm.

According to the New Zealand Pilot (11th edit., p. 234) tidal streams are weak in the bay but strong in the river entrances, setting northward with a rising tide and southward with a falling tide.

The chart for the area (N.Z. 6) published by the Hydrographic Branch, Navy Department, shows two tidal stations. Station A is located in the channel between Portland Island and Mahia Peninsula and Station B is about 15 miles south-south-east from Portland Island. The tidal streams at these stations are listed in table 1.

Table 1

Hours	A Lat. 39° 16′3S Long. 177° 52′1E			B Lat. 39° 31′9S Long. 178° 00′7E		
	Direction (Degrees)	Rate (Knots)		Direction (Degrees)	Rate (Knots)	
		Spring	Neap		Spring	Neap
6	241	1·6	1·3	198	0·2	0·2
5	247	1·3	1·0	174	0·3	0·2
4	251	0·8	0·7	160	0·3	0·2
3	323	0·1	0·1	142	0·3	0·2
2	046	0·8	0·7	120	0·3	0·2
1	047	1·3	1·1	090	0·2	0·2
H.W.	050	1·4	1·1	055	0·2	0·1
1	055	1·2	1·0	017	0·2	0·2
2	061	0·9	0·7	342	0·2	0·2
3	076	0·4	0·3	307	0·2	0·2
4	221	0·4	0·4	285	0·2	0·2
5	237	1·2	1·0	256	0·1	0·1
6	241	1·6	1·3	247	0·2	0·2

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Numerous rivers and streams discharge into the bay and very low salinities result (Garner, 1960). The country north of Hawke Bay is mountainous. From Wairoa to Tangoio Bluff (fig. 2) there are high cliffs backed by hills ending abruptly 12 miles north of Napier. From here to Cape Kidnappers

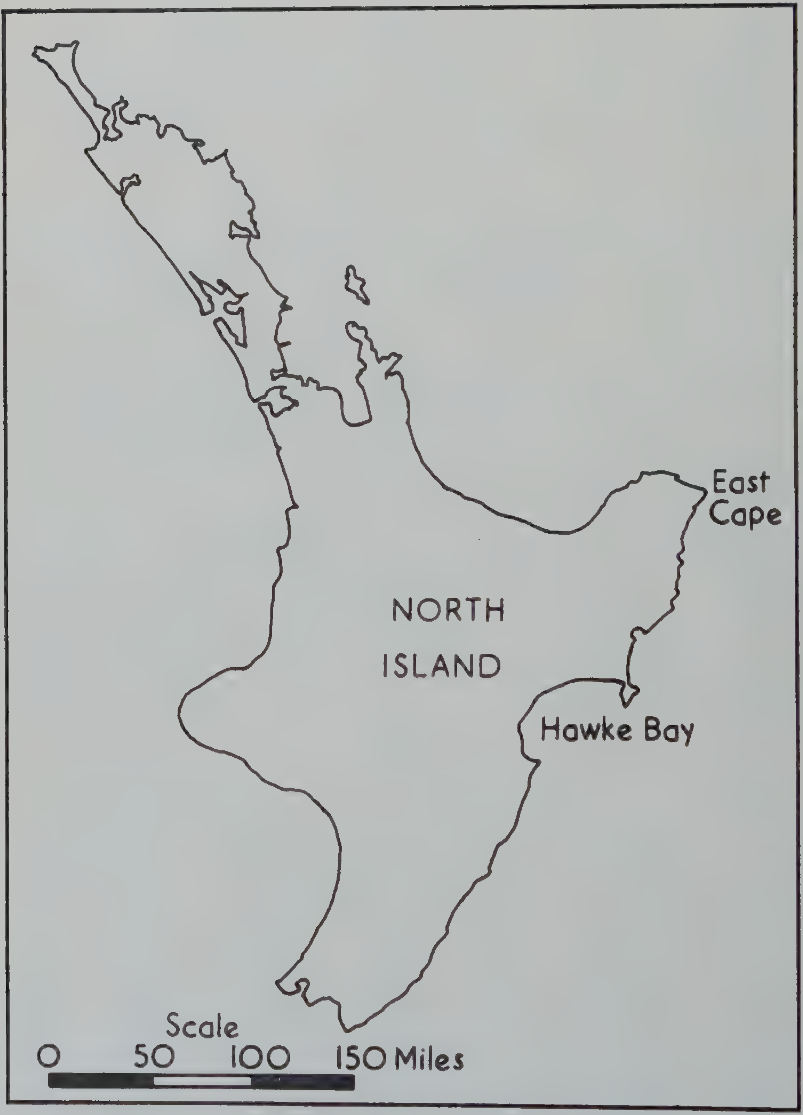

FIG. 1—Locality map showing Hawke Bay in relation to North Island.

extensive plains face the sea. The effect of the N.N.E. orientation of the land features upon the wind direction seems to be considerable; the variation in mean annual percentage frequency of wind direction at Portland Island, Wairoa, Napier, and Hastings is given by Garnier (1958) and shows a marked preponderance of winds from south-west and north-east for the southern stations, and a preponderance of winds from the north for the two northern stations.

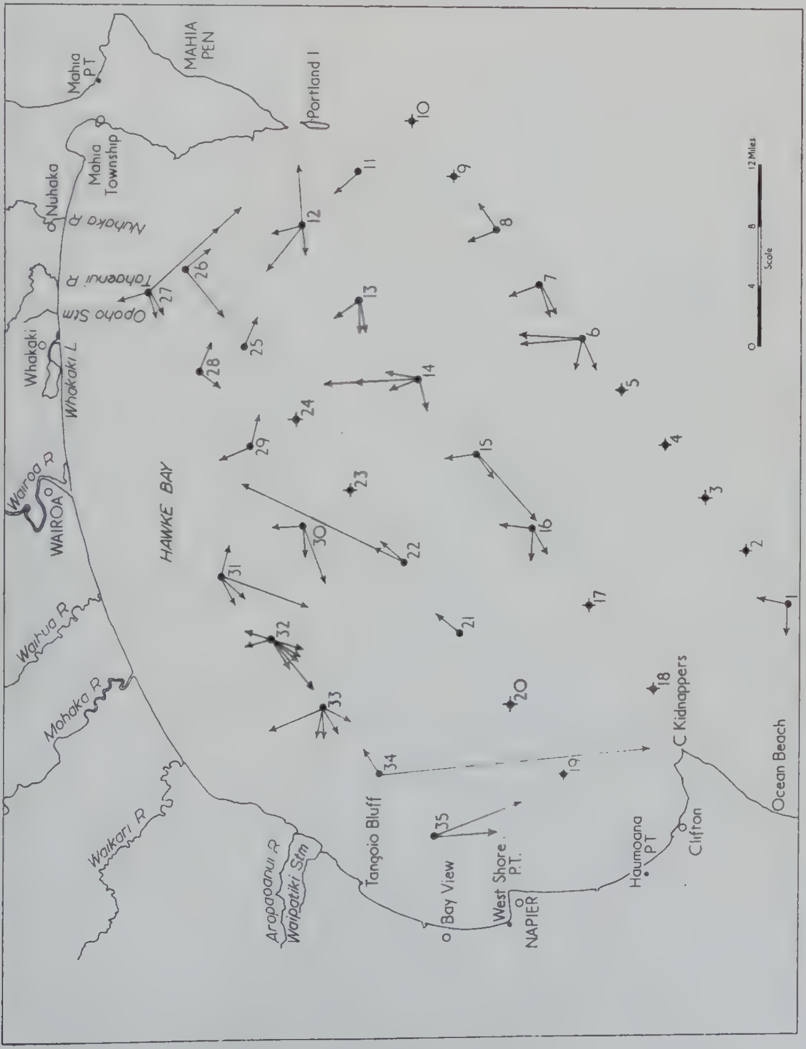

FIG. 2—Release points of cards. Arrows indicate direction of shortest tracks between release and recovery positions. Arrow lengths are proportional to the number of cards recovered in that direction.

The limited tidal information suggests that the tidal streams are oscillatory and are unlikely to contribute towards the establishment of a persistent surface circulation pattern.

DATA

Twenty cards were released from an aircraft over each of the 35 positions shown in fig. 3. The cards were released at an altitude of 1,000 ft and the wind during the period was north to north-west, 10–15 knots. The rate of fall

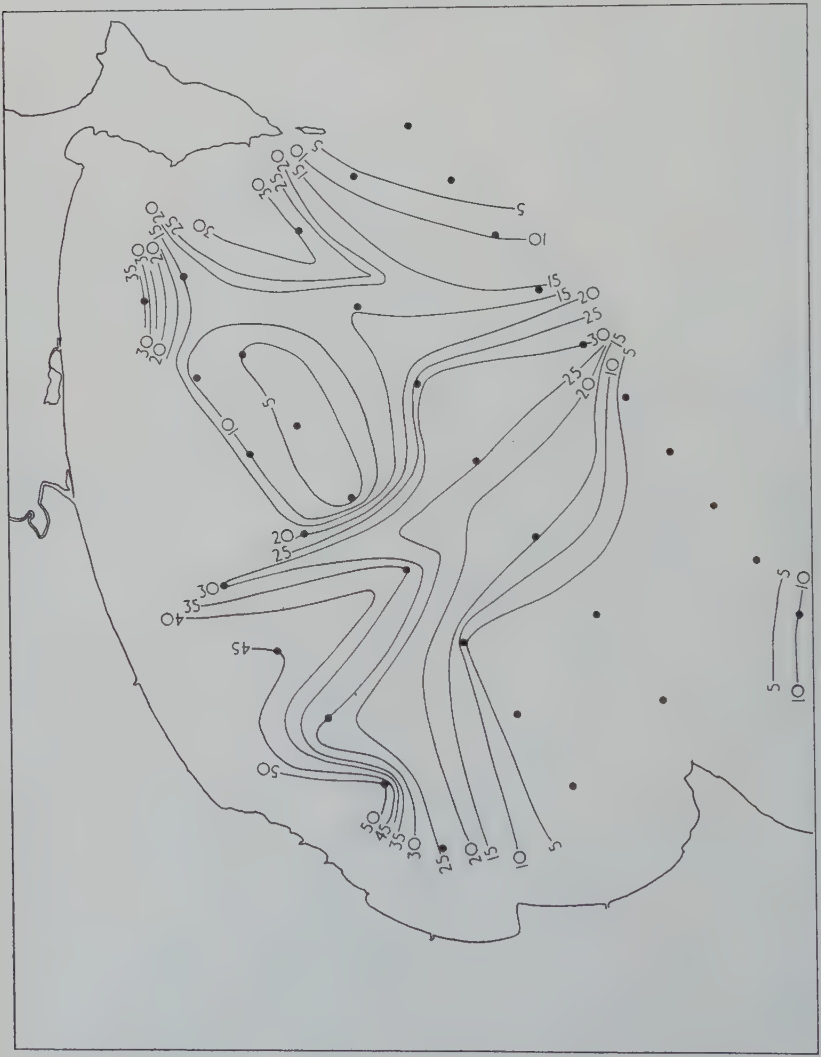

FIG. 3—Isolines based on the percentage recovery of cards from each release position.

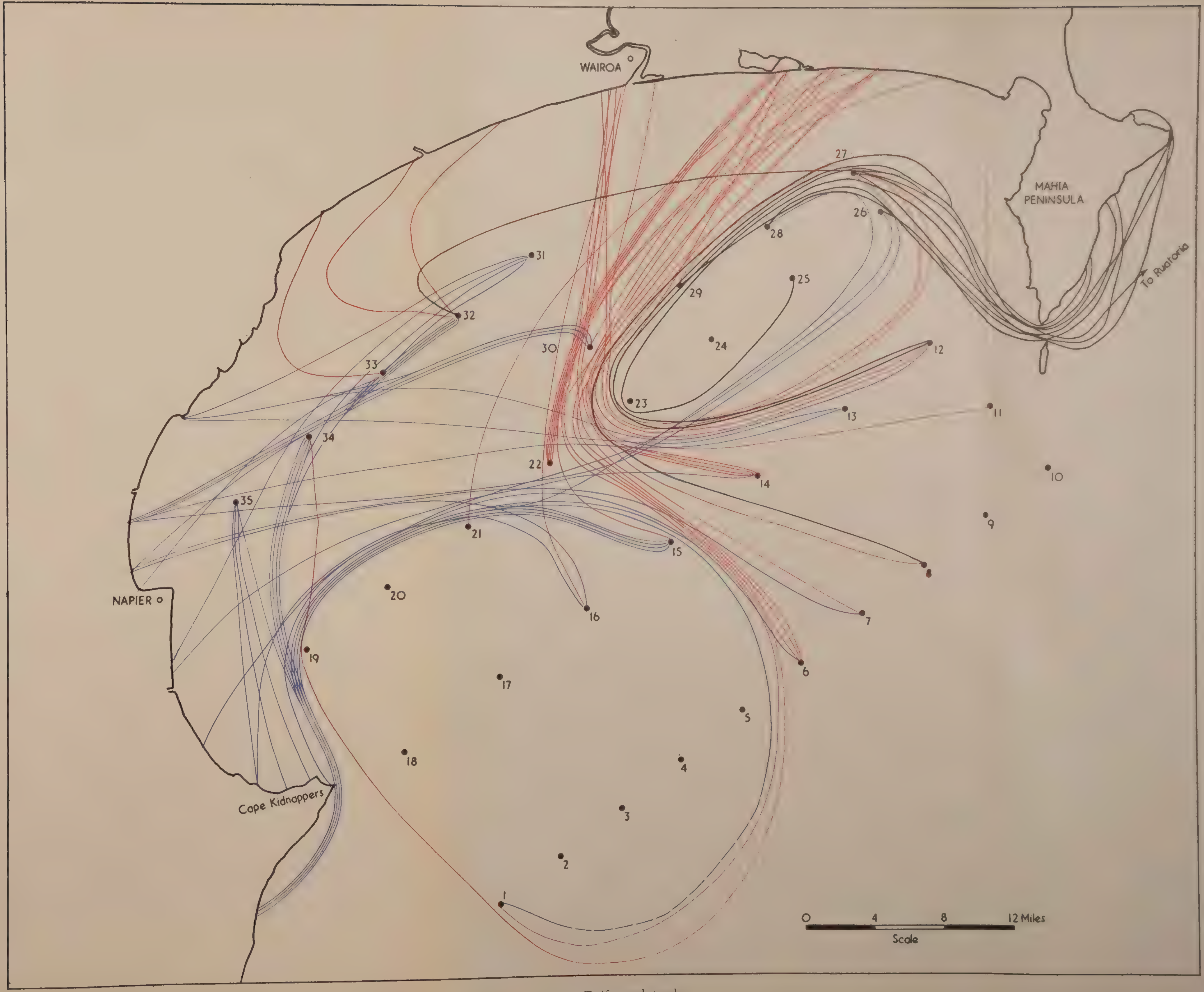

FIG. 4—Drift card tracks.

has been measured over a short distance at roughly 6½ ft per second. On these figures the cards released from the aircraft would be displaced ½ mile to ¾ mile in the direction of the wind. Ninety-nine cards have been returned. The percentage recovery of cards from each position has been plotted (fig. 3) and iso-percentage lines drawn. This percentage recovery pattern has been used as a basis for constructing drift tracks (fig. 4).

Wind speeds and directions for Hawke Bay for the period 13 December 1958 to 28 February 1959 were provided by the Meteorological Office, Wellington, and were taken from 3-hourly weather charts. A calendar of the recoveries up to 28 February 1959 is shown in fig. 5 and the wind data are incorporated in this calendar.

One of the difficulties in deducing drift rates from the interval of time elapsing between release and recovery is that cards could be stranded for a considerable time before being found. Some indication of this delay may be seen from the daily frequency of recoveries for each month (table 2), weekends as might be expected showing the highest returns. On one occasion in January, 5 cards were recovered by one person from one place and this has been treated as a single recovery in compiling the table.

TABLE 2—Daily Recoveries of Cards for Each Month Expressed as Percentages

	Dec.	Jan.	Feb.	Mar.	Apr.	May
Monday	18.2	6.5	19.0	0	0	0
Tuesday	0	8.1	9.5	0	0	0
Wednesday	0	9.7	19.0	0	0	0
Thursday	9.1	19.3	0	0	0	0
Friday	36.4	16.1	4.8	0	0	0
Saturday	0	11.3	9.5	0	0	50
Sunday	36.4	22.6	38.1	100	0	50
Total number of cards recovered	11	66	21	3	0	2

To indicate general directions of movement arrows have been drawn from each release position pointing directly towards the recovery points (fig. 2). The length of each arrow is proportional to the number of cards recovered in that direction.

DISCUSSION

Divergence of Cards

There is a considerable divergence in the movement of cards from most release positions (fig. 2). This divergence may occur either initially or at some time during the subsequent drifts of the cards. Since the cards were released at an altitude of 1,000 ft some dispersal would take place during their fall. No direct observation of the amount of this dispersal was possible but a rough estimate can be made. Twelve bundles of cards were released from a height of 20 ft and the maximum dispersion for each drop was measured; the mean of these measurements was 25 ft. From the "Theory of Random Walk," it may be expected that when the distance through which the cards fall is increased, the scatter will vary as the square root of the distance (Kennard, 1938). On this basis the maximum spread of cards released from 1,000 ft would be 175 ft. It can therefore be considered that the cards released at each position are sufficiently close together to be affected by the same drift-producing forces. Any divergence which occurs would then take place during the cards' subsequent drift.

Effect of Wind and Tides on Surface Drift

The wind data obtained from weather charts do not allow total wind-run figures to be compiled nor do the data show the variations which may exist over the bay. It would have been desirable to have measured the detailed airflow over the bay during the period of the experiment but facilities for making such measurements were not available. Hughes (1956) related gradient winds to the drift of plastic-enclosed cards and concluded that the water within 1 cm of the surface drifts on the average with 2·2% of the wind speed and in the same direction. The wind factor is reduced near coastlines where tidal streams are present (Bowden, 1953). No relationship between the winds for the period during which cards were afloat and the direction and rate of drift can be deduced from the data available, nor can a general relationship be deduced between the general wind pattern (fig. 5) and the drifts recorded during the three-month period. Some immediate effect of wind, limited perhaps to the actual strandings, can be seen from the calendar of recoveries (fig. 5); after south-east winds there were more recoveries than at other times.

Drift Tracks

Since the cards released at each position would be acted upon initially by the same drift-producing forces, the drift tracks in fig. 4 have been drawn to indicate this. Two areas of non-recovery are shown (fig. 3). It has been assumed that the surface movement in these areas is predominantly off-shore and the drift-tracks have been drawn to avoid passing through them although it is possible that some recovered cards have passed through non-recovery areas in their movement shorewards.

The experiment was widely publicised by J. Arthur Rank Film Distributors of N.Z. Ltd., who provided a reward for each card returned. The months of December, January, and February are those during which the coast is most frequently visited by holiday makers. A high recovery rate of cards was therefore to be expected and it is unlikely that many stranded cards were not returned.

The general pattern of surface drift (fig. 4) indicates a strong indraught into the centre of the bay, this indraught diverging to both north-east and south-west along the coastal margins of the bay.

Outward movement of water must therefore take place at the northern and southern ends of the bay. The recovery pattern provides no definite information regarding movement at the southern end, but it is clear that at the northern end cards moved out of the bay round the peninsula, and stranded on the coast nearby. One such card moved 100 miles north to the

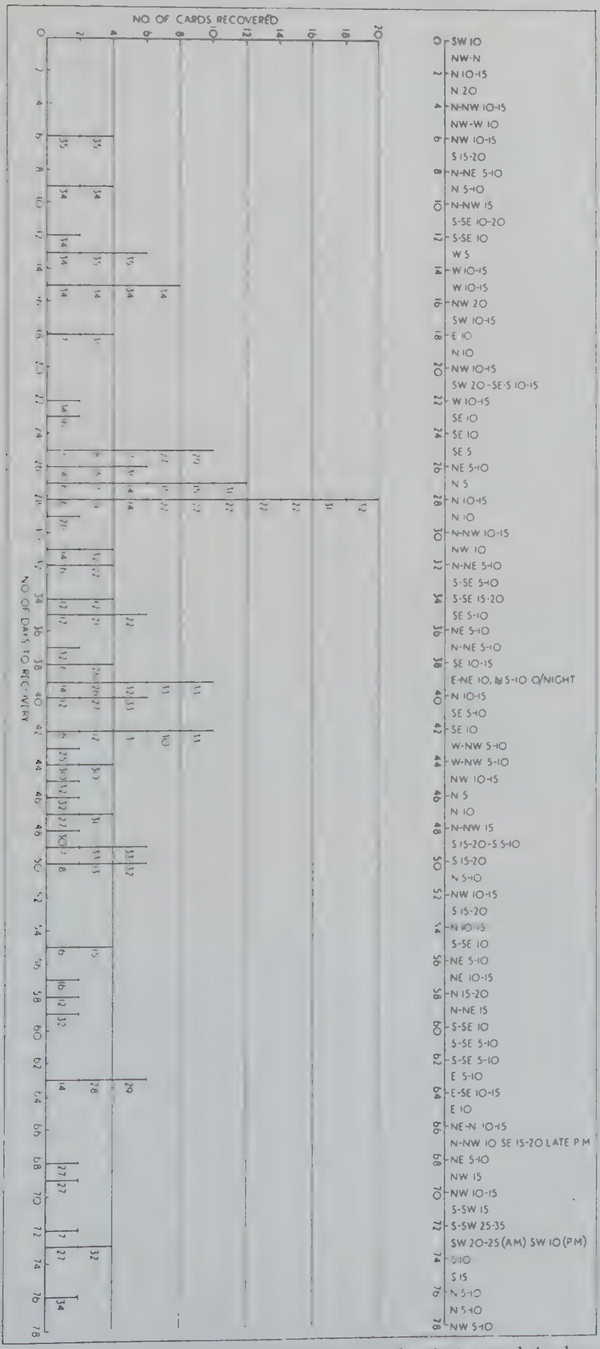

FIG. 5—Calendar of recoveries. Wind directions and wind speed in knots are shown along the top.

vicinity of East Cape. Fleming (1950) and Garner (1953) have demonstrated the existence of a permanent south-flowing current, the East Cape Current, off the east coast of North Island, whilst Brodie (1960) has suggested that the north-flowing Canterbury Current can be present further inshore as far north as Gisborne. It is likely that in the present instance the outflowing current at the north end of the bay was affected by the northward flowing Canterbury Current, which probably exerts control over the entire Hawke Bay circulation as postulated in this paper.

Persistence of Current Pattern

The effect of the offshore current as a controlling factor in setting up the circulation pattern is liable to fluctuation as the boundaries between the southflowing East Cape Current and the northflowing Canterbury Current themselves fluctuate (Garner 1959).

The persistence of the current pattern suggested in this paper (fig. 4) cannot be determined from a single drop of cards. However, the results obtained are broadly consistent over a three-month period.

CONCLUSION

The surface water movements in Hawke Bay as derived from the recoveries of drift cards reveal a simple circulatory pattern. The main inflow takes place approximately along the mid-line of the bay. This current bifurcates and the two currents thus formed follow the coast line and leave the bay at the northern and southern extremities. This surface water then appears in the present instance to move northwards with the Canterbury Current. These water movements do not appear to be directly related to the winds over the bay but are probably controlled by the coastal current system.

ACKNOWLEDGMENTS

These investigations were carried out in cooperation with the J. Arthur Rank Film Distributors of N.Z. Ltd., who underwrote most of the expenses and offered suitable rewards to those who returned cards. Their interest and assistance, particularly of their General Manager, Mr P. J. Croft, is gratefully acknowledged. Thanks are also due to Mr A. E. Gilmour who released the cards; Dr J. F. Gabites for meteorological data; Miss J. A. Leece who helped in acknowledging and recording cards; and to the members of the public who returned cards.

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INFAULTED TRIASSIC OUTLIER AT LAKE FERGUS, EGLINTON VALLEY, SOUTHLAND

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(Received for publication, 21 July 1959)

Summary

Triassic sediments are preserved in a narrow infaulted outlier at Lake Fergus on the western side of the Hollyford Fault. Three fossil species show that the beds are Otamitan.

INTRODUCTION

In 1954 Messrs A. R. Mutch, B. L. Wood, and G. W. Grindley discovered fossils in boulders of indurated, dark muddy sandstone in a creek bed west of Lake Fergus (N.Z.G.S. 6043). The fossils were tentatively referred to *Atomodesma* and *Pseudomonotis* in Grindley, (1958, p. 43), and seemed to indicate that the beds from which the boulders had been derived were of Permian age. The beds were mapped on a 4-mile reconnaisance map by Grindley (1958) as belonging to the Howden Formation, lowest formation of the Maitai Group in the Eglinton Valley region.

OTAMITAN FOSSILS

Further examination of the fauna showed that an unbonal fragment mistaken for *Atomodesma* belonged to the Triassic genus *Mysidioptera* Salomon, and a second visit to the fossil locality by Messrs A. R. Mutch, P. Chandler, and J. B. Waterhouse in 1958 yielded good specimens of *Mytilus problematicus*. Three lamellibranchs are now known from GS 6043:

Mysidioptera.
"Mytilus" problematicus Zittel.
? Halobia.

All three indicate an Otamitan (Carnian) age. *Mysidioptera* Salomon, not previously recognised in New Zealand, is represented by a widespread non-costate species in the Otamitan of the North and South Islands: the Eglinton specimens are conspecific with this species. *"Myrtilus" problematicus* is the key fossil of the Otamitan Stage. Several moderately inflated right and left valves are available, but none are as high as the inflated form named *mirabilis* by Trechmann (1918, p. 202). Crushed fragments of a pteriid named *Pseudomonotis* in Grindley (1958) possibly belong to *Halobia.*

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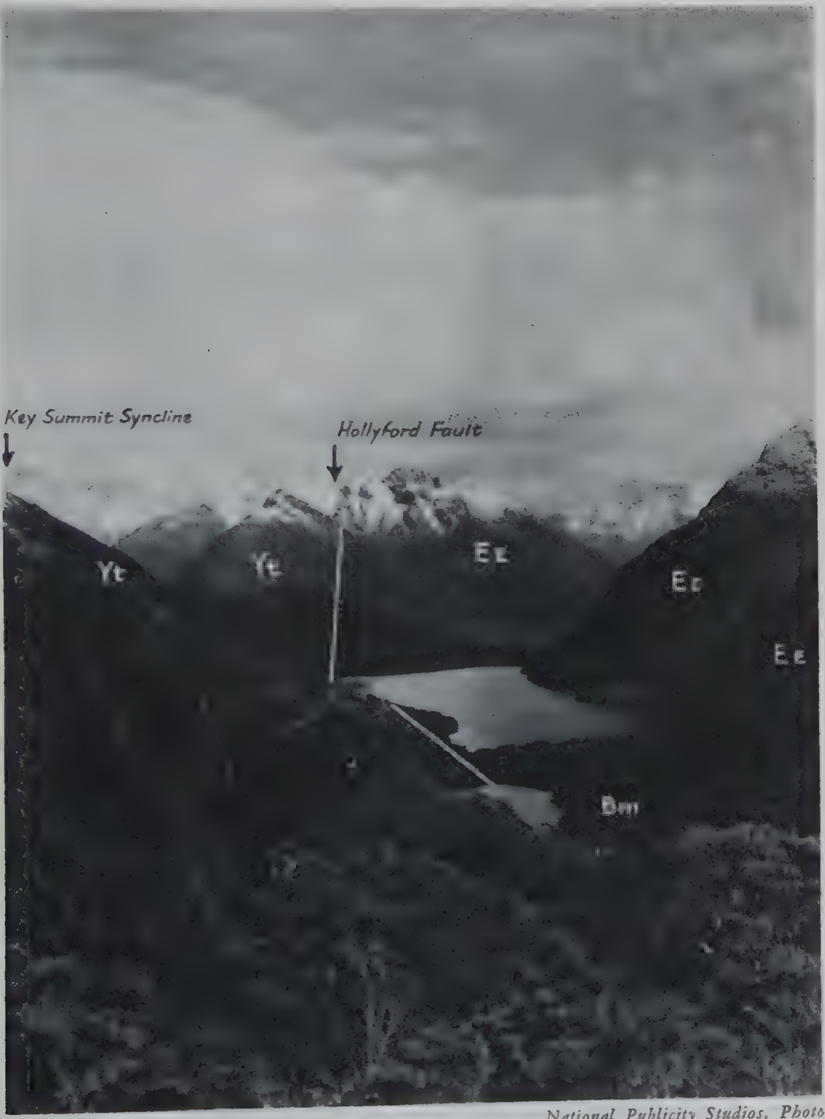
REVISED STRUCTURE

Previously these Otamitan beds were mapped as Maitai immediately to the east of the Hollyford Fault – a major fault which passes along the upper



National Publicity Studios, Photo.

FIG. 1—Upper Eglinton Valley and Livingstone Range looking south from near Key Summit. Hollyford Fault separates Maitai argillites and sandstones (Yt) on left from Te Anau volcanics (Ee) on right. Upper Triassic beds (Bm) downfaulted on west side of Lake Fergus (front).

Eglinton Valley (fig. 1) into the Hollyford Valley, separating Maitai sediments from the older Te Anau (Eglinton) volcanics to the west. (Grindley 1958, map 2). The revision of the paleontology suggests that the Hollyford Fault lies east of the fossiliferous beds. This correction brings the structure much more into agreement with that of the Permian and Triassic belt at Nelson, where a narrow synclinal strip of Triassic beds is separated by the Eighty-eight Fault (the Nelson counterpart of the Hollyford Fault) from a synclinal fold in the Maitai beds, comparable with the Key Summit syncline in the Maitai beds of the Eglinton and Hollyford Valleys.

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RECENT VARIATIONS OF CHLORIDE CONTENT AND SPRING ACTIVITY AT WAIRAKEI GEYSER VALLEY

By G. E. K. THOMPSON, Geophysics Division, Department of Scientific and Industrial Research, Wellington

(Received for publication, 16 November 1959)

Summary

Repeat measurements of chloride, electrical conductivity and pH of samples from 11 selected springs in Wairakei Geyser Valley have indicated a downward trend in both chloride content and electrical conductivity since the measurements were initiated in 1953. Changes of pH have been irregular. The recorded effects are shown to be consistent with a decrease in the amount of primary chloride water feeding the area.

INTRODUCTION

In 1950-51 Gregg and Laing (1951) located and mapped all the springs in Wairakei Geyser Valley, and in the course of their work took samples for the determination of chloride content, electrical conductivity and pH. Because these measurements provide useful information on the nature of the feeding sources, they were repeated from time to time by the author. In a previous paper (Thompson, 1957), it was shown that in the two-year interval following the measurements of Gregg and Laing, a small decrease in the chloride content of the discharges had occurred, and this paper is intended to bring the comparison up to date.

MEASUREMENTS

The results of chloride determination, electrical conductivity and pH measurements have been set out in tables 1, 2, and 3 respectively. The location of each sampled spring can be seen in fig. 1 and the chloride variations are shown graphically in fig. 2.

From table 1 it is apparent that there have been large decreases in chloride content in all the springs sampled with the exception of spring 190 (Dancing Rock Geyser) and possibly spring 97 (Champagne Pool). The average percentage decrease on the 1951 figures is 54%, the maximum being 99% in spring 113 which ceased to discharge in January 1954.

From table 2 it will be seen once again that apart from the Dancing Rock Geyser (spring 190), there is a general decrease in measured values since the 1951 measurements. Percentage decreases average 50% with the maximum decrease (83%) found in spring 174. Because the electrical conductivity of the Wairakei Geyser Valley springs is mainly due to the chloride ion, these decreases in conductivity are not unexpected.

The results in table 3 show no general trend. However, in the case of spring 38 (Dragon's Mouth Geyser) and spring 18, there is a trend toward greater alkalinity of the sample, whereas spring 197 and spring 115 show the opposite trend.

TABLE 1—Chloride Content (parts per million)

Spring No.	1951	1953	1955	Feb. 1956	Mar. 1957	Aug. 1957	Nov. 1958	Jan. 1959	Sept. 1959
198 Ocean Geyser	1470	1084	874	1121	950	780	......	......	638
197 Rainbow Pool......	1531	1424	1316	1379	1120	1140	894	908	837
201	1426	1322	1114	1184	870	960	......	......	659
190 Dancing Rock Geyser	1728	1768	1754	1753	1720	1620	1716	1723	1702
174	750	250	158	184	112	126	......	......	85
113 Opal Pool	1624	1536	1465	1634	100	......	......	......	14
38 Dragon's Mouth Geyser	1860	1806	1114	1042	770	720	......	......	496
18	1803	1606	1174	1184	1020	980	......	......	766
55 Waitangi Pool	1556	866	1025	1031	1010	895	......	......	766
59 Great Wairakei Geyser	1709	1082	1096	1074	1030	930	......	......	837
97 Champagne Pool	1739	1674	......	...	1560	1640	1532	1553	1560

NOTE—The exact month of the early samplings are unknown.

TABLE 2—Electrical Conductivity (in Units of 10^{-5} Reciprocal Ohm Centimetres)

Spring No.	1951	1953	1954	1955	Feb. 1956	Aug. 1957	Sept. 1959
198 Ocean Geyser	342	313	304	372	242	......	208
197 Rainbow Pool	411	395	376	378	402	......	254
201	378	372	376	335	333	......	215
190 Dancing Rock Geyser	455	421	476	506	495	440	505
174	218	89	89	70	71	......	37
113 Opal Pool	466	450	392	442	490	......	101
38 Dragon's Mouth Geyser	506	489	446	338	320	215	174
18	482	444	397	350	355	273	252
55 Waitangi Pool	436	262	276	320	314	283	215
59 Great Wairakei Geyser	450	313	318	328	314	283	263
97 Champagne Pool	479	464	472	......	......	440	445

NOTE—The exact month of the early samplings are unknown.

TABLE 3—pH Determinations

Spring No.	1951	1953	1954	1955	Feb. 1956	Aug. 1957	Sept. 1959
198 Ocean Geyser	6.2	6.7	6.0	4.9	4.6	4.7	5.9
197 Rainbow Pool	6.7	6.7	6.5	5.3	4.5	4.5	3.9
201	7.4	7.4	7.5	6.9	7.3	7.0	6.8
190 Dancing Rock Geyser	7.6	8.0	8.0	8.0	8.8(?)	7.7	8.0
174	7.2	7.2	7.2	6.9	7.8	8.0	8.7
113 Opal Pool	6.0	3.5	3.4	4.9	4.0	......	......
38 Dragon's Mouth Geyser	8.2	8.6	8.8	8.9	8.8	8.8	9.3
18	8.3	8.6	8.7	8.8	8.5	8.6	9.1
55 Waitangi Pool	7.1	7.3	8.0	6.8	7.3	6.2	6.1
59 Great Wairakei Geyser	7.6	8.1	7.9	7.6	7.5	7.0	7.4
97 Champagne Pool	7.6	7.1	7.7	......	......	7.4	7.5

NOTE—The exact month of the early samplings are unknown.

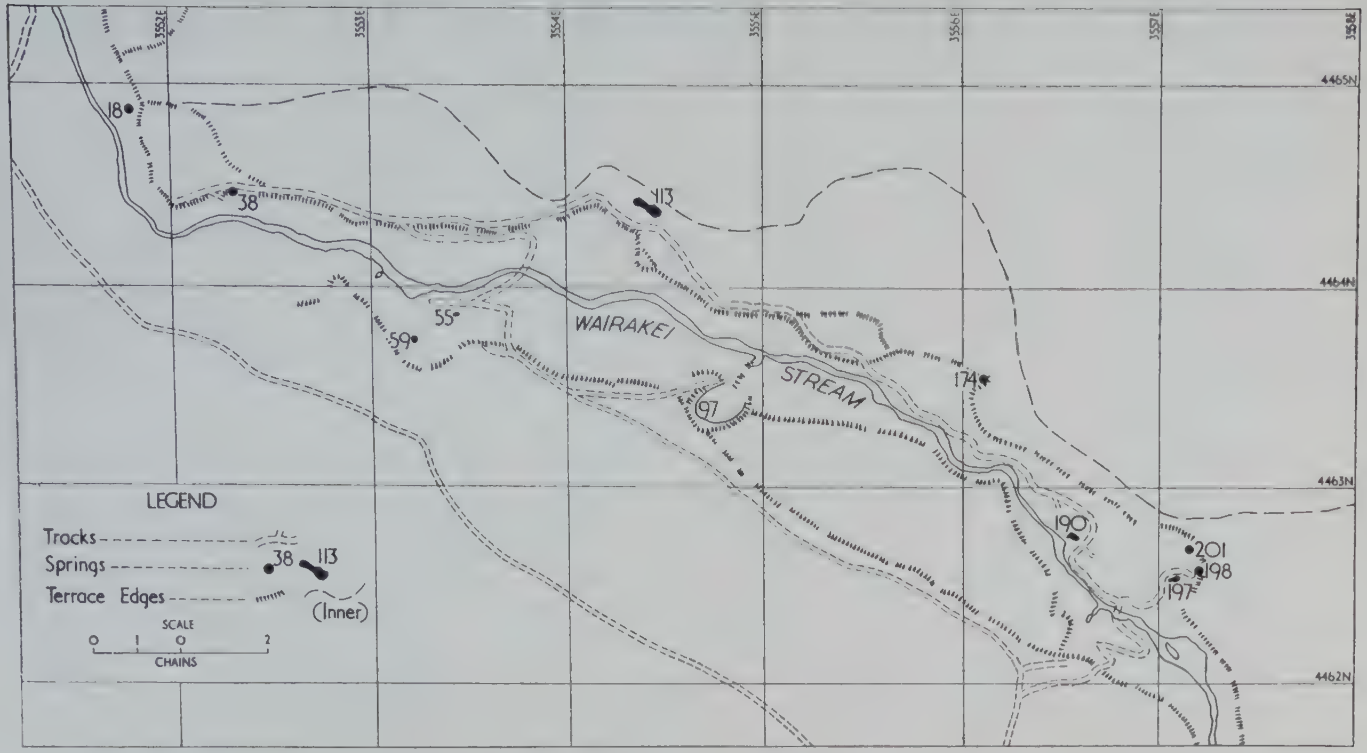

FIG. 1—Wairakei Geyser Valley hot springs.

Discussion

In making the original choice of the springs to be measured, an attempt was made to obtain as wide a coverage as possible, having regard to the elevation and characteristics of the springs concerned. For example, spring 18 was a vigorously boiling, slightly superheated pool, situated 5 m from and about 3 m above the level of the Wairakei Stream, while spring 174 had a temperature of 77°C and was located in a narrow gully approximately 9 m above stream level and some 50 m from it. However, there has been considerable change in the nature of the spring discharges since measurements first commenced in 1953. In September 1956 spring 18 ceased to discharge; in May 1954 the last recorded eruption of the Great Wairakei Geyser (spring 59) took place; and no eruption of the Dragon's Mouth Geyser (spring 38) has been observed since the beginning of 1958. Discharge curves for springs 18, 113, and 174 are shown in fig. 3.

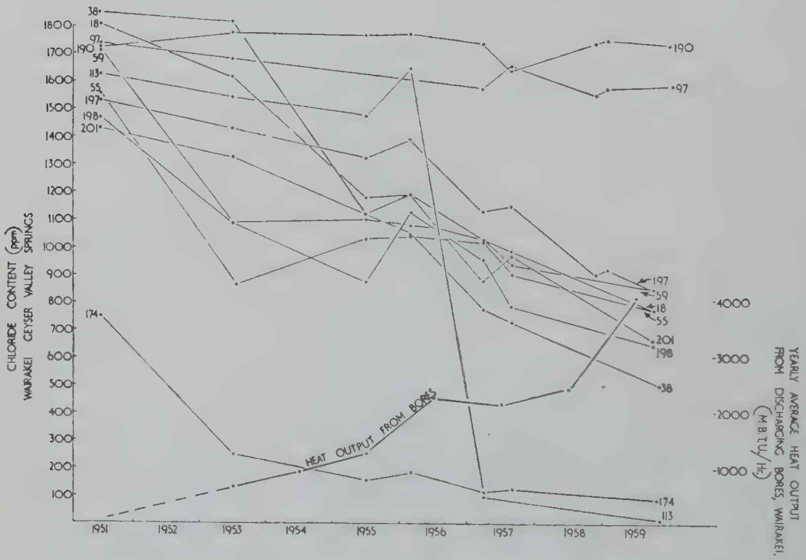

Fig. 2—Chloride Content of Geyser Valley hot springs and heat output from boreholes.

There are several phenomena which might conceivably produce the changes observed over the years. Firstly, if it is supposed that the discharge from springs consists of some primary chloride water together with a proportion of ground water, the effects observed would be accounted for by postulating the addition of extra ground water, which would contain little chloride. However, if this were the case, one would expect an increase in the mass discharge from the springs, and this has certainly not occurred. Alternatively, the decrease in chloride could be accounted for by postulating the replacement of primary chloride water by free steam, but this possibility must be ruled out for temperatures generally have fallen. A third possibility is that the

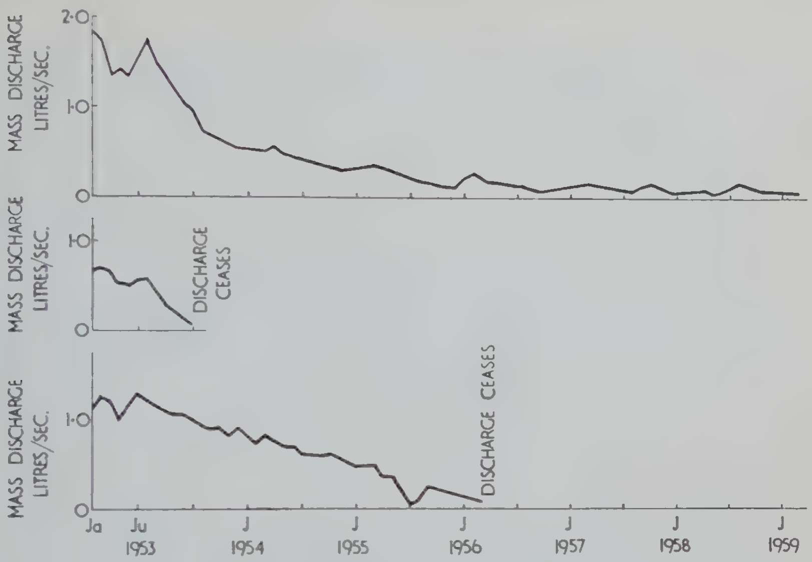

FIG. 3—Mass discharge variations.

amount of primary chloride water in the original discharge has decreased. This hypothesis is confirmed by the fact that springs at higher elevation were the first to show a decrease in mass discharge, while those close to the level of the Wairakei Stream were not affected until later. For example, spring 113 ceased to discharge in January 1954, but it was not until September 1956 that spring 18 was similarly affected. However, it would be unwise to generalise as when the discharge ceased in spring 113 its temperature rose from 85°C to over 95°C. where as the reverse held at spring 18 where the temperature dropped from boiling point to around 86°C.

Whatever the explanation, there is no doubt that the activity of springs at Wairakei Geyser Valley has declined over the years and no new activity has been reported to make up for the decrease. Measurement of chloride content appears to be a good indicator of trends in activity.

The question that immediately comes to mind is whether the intensive drilling programme being carried out by the Ministry of Works in the vicinity (D Line Wairakei) could be responsible for this change. It is a fact that large amounts of heat are being withdrawn from the country by discharging bores and fig. 2 shows how this drawoff has increased over the years. An inverse correlation between chloride content of spring discharges and heat output from the bores clearly does exist, but it must be remembered that the existence of large deposits of sinter in areas that are no longer active suggests that the thermal areas are subject to natural fluctuations. There are indications that in other parts of the Wairakei field the activity may actually have increased in the period under observation.

ACKNOWLEDGMENT

Since 1957, measurements of chloride content and pH have been made by the Dominion Laboratory staff at Wairakei, and the results are published here with their permission.

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MELLARIUM, A NEW TRIASSIC PLEUROTOMARIID GASTROPOD FROM NEW ZEALAND

By J. B. WATERHOUSE, New Zealand Geological Survey, Department of Scientific and Industrial Research

(Received for publication, 23 June 1959)

Summary

A new pleurotomariid *Mellarium* gen nov. occurs in the Anisian (Middle Triassic) of the Southland Syncline. Superficially the new genus is similar to *Shansiella* Yin, a late Paleozoic gastropod. Two new species of *Mellarium* are described, *M. mutchi*, from lower Etalian greywacke, and underlying breccias, and *M. nodulosum*, from the upper Etalian. A crushed gastropod from the upper Countess Formation of the Eglinton Valley is possibly conspecific with *M. mutchi*

INTRODUCTION

A relatively common fossil in the pre-Kaihikuan Middle Triassic rocks of the Southland Syncline is a globular turbiniform pleurotomariid with spiral costae or nodules. In some respects the shells are similar to *Shansiella* Yin (1932) from the Carboniferous and Permian of the Northern Hemisphere, but the Triassic species of Southland warrant generic separation, because they differ consistently in some details of the slit, selenizone and whorl-profile. No described Triassic genus seems to be closely related to the New Zealand shells.

Two species are known; the types of both are deposited at the N.Z. Geological Survey, Lower Hutt, unless otherwise stated. Terminology used in the gastropod descriptions follows that of Cox (1955).

SYSTEMATIC DESCRIPTION

Genus *Mellarium* new genus

(Latin – *mellarium* - neuter, meaning a bee-hive, from the fancied resemblance of these globular shells with spiral ornament to the old-fashioned bee-hive.)

TYPE SPECIES. *Mellarium mutchi* n. sp.

DIAGNOSIS. Cryptomphalous, medium-sized globular turbiniform pleurotomariids; slit placed above mid-height of the whorl, one fifth of the whorl circumference deep; selenizone narrow and convex, either smooth or ornamented with nodules; whorl periphery lies below selenizone, whorl profile subangular, and slightly concave above the slit, base flattened; whorls ornamented by strong revolving costae or nodules.

DISCUSSION. The upper Paleozoic genus *Shansiella* Yin (1932; Knight, 1941, p. 317) closely resembles the New Zealand genus in shape, cryptom-

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phalous base, and ornament, but the aperture differs in several aspects. In the New Zealand shells the upper and particularly the lower apertural margins are convex towards the aperture. Where described in *Shansiella* the margins are straighter, apart from minor sinuosity, as in the Scottish *S. globosa* (Thomas) and *S. globosa nitida* (Thomas), of which the types have been examined at the Royal Scottish Museum, Edinburgh. The Scottish species were originally placed in *Latischisma* Thomas, but this name was synonymised with *Shansiella* by Knight (1944, p. 457). The lower apertural margin is also straight in the American species *S. planicostata,* according to Girty (1937). The slit lies close to the upper third of the whorl profile, and extends for one-fifth of the whorl circumference in the two species of *Mellarium,* whereas the slit lies below mid-height of the whorl in *Shansiella,* and is just over one-sixth of the whorl circumference in *S. globosa nitida* (Thomas), the one species of *Shansiella* in which the depth of the slit has been described.

In *Mellarium* the selenizone is comparatively narrow, convex, and without spiral ornament. That of *Shansiella* is broad and generally concave. Two to five spiral threads are generally present on the selenizone of the Scottish *Shansiella,* and are also reported by Dorsman (1945, p. 60) in *S.* cf. *carbonaria* (Norwood and Pratten 1855) from the Westphalian C of the Netherlands. It is unknown if spiral threads are present on the selenizone of other species of *Shansiella.*

The upper whorl profile of *Mellarium* has a concave zone immediately above the selenizone; this is lacking from *Shansiella.* The base of *Shansiella* is more rounded, except in the American species, *S. planicostata,* and specimens compared with this species by Dorsman (1945) and Frech (1911).

Mellarium mutchi n. sp. figs 3–12

Named after Mr A. R. Mutch, N.Z. Geological Survey, Dunedin.

MATERIAL. One immature and ten mature specimens are available from GS 6066. Most features are preserved, in spite of the incompleteness of any one shell, and the depth of the slit is known in five shells. Over twenty fragmentary or decorticated specimens and external casts are also present from GS 6066 and 6326. Poor external and obscure internal casts occur at GS 5199, and single internal casts with fragments of the exterior are known about 1,000 ft stratigraphically above GS 6066 and 6323, at GS 3857 and 3593. Fragments of the internal and external cast of the body whorl are available from the northern limb of the Southland Syncline, at fossil locality S 179/655. (See table 3.)

In order to obtain fine details of the external ornament, several specimens were leached in dilute hydrochloric acid, and moulds were prepared from the resulting casts with polyvinyl chloride in dibutyl pthalate. Other specimens were sectioned to study the columella, and shell structure.

TYPES:

HOLOTYPE: Specimen T.M. 2040, GS 6066; figs. 3–5.

PARATYPES: Specimens T.M. 2041–2052, GS 6066.

TABLE 1. Dimensions (approximate in millimetres) of Specimens of *Mellarium mutchii* from GS 6066.

Specimen TM	Width W	Height H	Height Body Whorl B	H/ /W	B/ /W	Width Aperture	Pleural Angle
2052	19	19		1·0		10	115
2050	26	29	18	1·12	0·62	12	87
2042	28	24	16	0·86	0·66	14	115
2043	29	? 26	17	0·71	0·65	8	105
2045	29	? 27	18	0·93	0·68	15	110
2047	29	29	16	1·0	0·55	14	95
2046	30	30	17	1·0	0·57	12	95
2048	31	26	17	0·83	0·66	15	118
2041	31	28	19	0·91	0·68	17	110
2040	32	29·5	18	0·92	0·62	15	95
2051	33	31·5	23	0·95	0·72	18	120
2044	35	34	19·5	0·98	0·58	18	95

DESCRIPTION. Comparatively large for the genus, the shells are globular turbiniform pleurotomarians, ornamented by strong costae. The nucleus is unknown: the succeeding whorl is depressed turbinate, and without ornament. Below are four whorls that increase progressively in size, approximately by a factor of two, each whorl embracing the preceding whorl up to the base of the slit-band. Sutures are filled in with callus.

The whorl profile is moderately rhomboidal, with a sutural shoulder, a weak angulation at the selenizone, and a stronger angulation, usually peripheral, above the flat cryptomphalous base. The profile is slightly concave immediately above the selenizone.

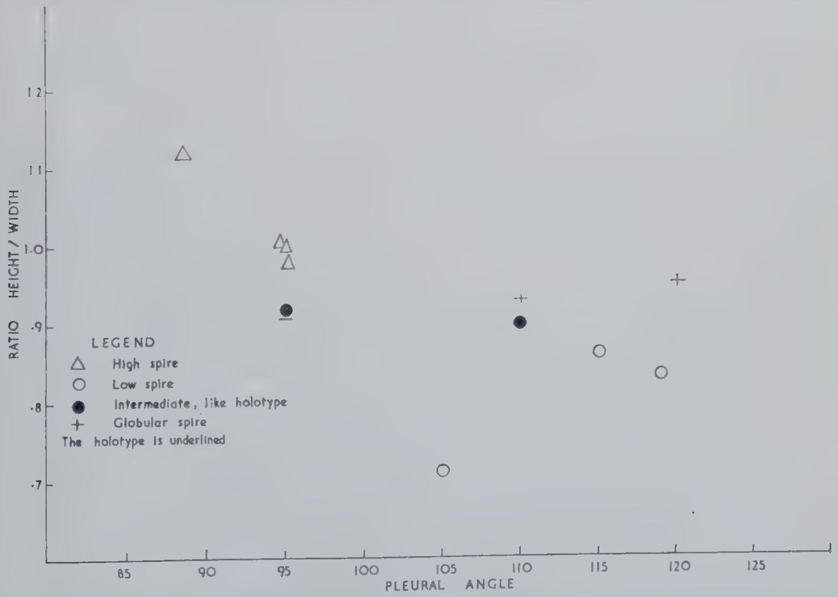


FIG. 1—Graph of the height/width ratio plotted against pleural angle, distinguishing high-spired from low-spired shells. Some globular shells with relatively high spire have a low pleural angle.

The lower part of the inner lip of the aperture is strongly thickened into an arched columella, of which the base extends halfway across the basal lip. No parietal thickening is developed. As a whole the aperture slopes prosoclinally at about 45° from the suture. Above the selenizone the margin is prosocline, lying at 60° to 70° from the suture, with the maximum convexity at the sutural shoulder. Between the selenizone and the lower angulation the margin arches towards the aperture, leaving the slit at an angle of 20° to 30°, and then turning through about 90°. Below the angulation the apertural margin is prosocline, and sinuate in outline, being concave towards the aperture near the angulation, and becoming gently convex or straight as it approaches the columella.

The depth of the slit is one-fifth of the circumference. The selenizone is gently convex and bordered above and below by a narrow groove, with a narrow lira each side in some specimens. Apart from closely spaced lunulae the selenizone is smooth.

The shell is ornamented by strong spiral costae separated by wide concave interspaces. A somewhat granular appearance is imparted on some specimens to the costae by the growth-lines, which divide the costae into low nodules. Four primary costae lie above the selenizone, with one or two secondary costae on the shoulder, and one or two below. Three, or rarely four, primary costae lie between the selenizone and the basal angulation, and a secondary costa may occur in one or other interspace. The second costa below the selenizone is the strongest, and in some shells lies at the periphery. Below are six to eight subequal basal costae, with few or no secondary costae. The number of primary costae is much the same in the penultimate whorl, but secondary costae are usually absent.

The shell is 2 mm thick, and comprises a thick median layer of recrystallised calcite, with a thin black layer on each side, secondarily composed of chlorite.

VARIATION. Specimens from the type locality GS 6066 vary in whorl shape and in the pleural angle. The holotype and another well preserved specimen are intermediate in character between shells with a low pleural angle, and shells with a high pleural angle (figs. 1, 2). The low-spired shells have a comparatively rounded whorl and more horizontal base, whereas the high-spired shells have a more elongated whorl section with a prominent peripheral angulation and a more steeply inclined base. Another form is represented by specimens with unusually large and globular whorls.

Shells from younger beds at GS 5199, 3593, 3587 and 6326 are small, and too badly preserved for adequate comparison, but appear to have coarse costae like the types from GS 6066.

Turbinate cryptomphalous specimens from the northern limb of the Southland Syncline at Kaka Point are probably conspecific with *M. mutchi*. A fragment of the external cast of a body whorl CG 8, kept at the Geology Department, University of Otago, shows the convex selenizone and the ornament of strong spiral costae typical of the species, and the whorl profile matches that of the somewhat low-spired paratypes. Judging from associated fragments, the growth lines trace a similar course, but the spiral costae above the selenizone are lower than in the paratypes.

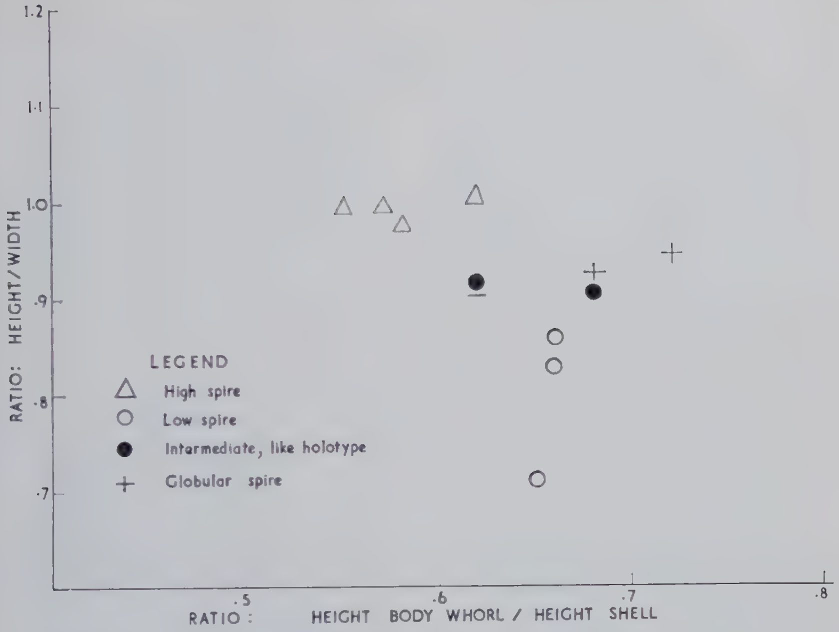


FIG. 2—Graph of the height/width ratio against the ratio of the height of the body whorl to the height of the shell. The shells range through high-spired forms with a relatively high body whorl to low-spired forms with a low body whorl. The globular shells are intermediate in position.

RESEMBLANCES. Although *Mellarium* differs generically from *Shansiella*, species of both genera are superficially similar. Like *M. mutchi*, *S. altispiralis* is a large shell with few strong costae, but has a higher spire, more rounded base, and slightly more numerous costae. *S. carbonaria* (Norwood and Pratten) has a more angular profile with a very well defined basal angulation and a somewhat similar flat base. Costae are more numerous.

AGE. *M. mutchi* is Anisian. In the Etalian beds at GS 5199 doubtful specimens are associated with *Leiophyllites marshalli* Browne, and species of *Beaumontites* Browne (1952), which has been synonymised with the widespread Anisian genus *Parapopanceras* by Kummel (1953, p. 16). An ammonite at GS 6326 also indicates a middle Triassic age (Dr B. Kummel, letter, 22 October 1956). Together with the underlying beds at GS 6066, the beds at GS 6326 belong either to the lower part of the Etalian Stage, or to a new sub-Etalian stage, as yet undefined (Mr A. R. Mutch, pers. comm.).

The specimens from Kaka Point occur about 3,200 ft stratigraphically below a Kaihikuan fauna, and a few hundred feet below Anisian ammonites identified by Dr B. Kummel (Mr J. D. Campbell, University of Otago, pers. comm.).

? ***Mellarium*** cf. ***mutchi*** Waterhouse. figs. 13, 14.

1958 ? *Shansiella* n. sp. Waterhouse in Grindley: 47.

MATERIAL. One specimen is available from greywacke grit of the upper Countess Formation, GS 4514.

Dimensions (in millimetres) of crushed specimen of *Mellarium* cf. *mutchi*

Specimen	Width	Height	Width Aperture
TM 2054	36	24	18

DESCRIPTION AND RESEMBLANCES. The specimen is a globular turbinate, badly crushed specimen with a crytomphalous flat base and strong revolving costae. Although the columella is more prominent than in *M. mutchi*, its height has been exaggerated by decortication from the adjoining base. Growth-lines indicate that the aperture resembles that of *M. mutchi*, but unfortunately the slit and selenizone are largely unknown. The costae are comparatively fine, but may lie within the range of variation for *M. mutchi*.

The specimen is possibly conspecific with *Mellarium mutchi*, but crushing has so distorted the specimen that specific and generic comparison is hazardous.

AGE. The age of the Countess Formation is uncertain. In the Geological Map of New Zealand (1958) and in the Eglinton bulletin, Grindley (1958) suggests that the Countess Formation is Upper Permian. However the higher part is possibly lower or middle Triassic, because the gastropod from it can be tentatively matched with a Triassic species.

Mellarium nodulosum n. sp. figs. 15–22.

MATERIAL. Five specimens are available from the upper Etalian fine greywacke mudstones of the south limb of the Southland Syncline. Three are partly decorticated specimens from GS 5519. The other two, from GS 6661 and 6075, were collected with the surrounding matrix, and have been leached in hydrochloric acid, to prepare polyvinyl chloride moulds from the external casts. The nucleus is obscure, and the outer aperture destroyed.

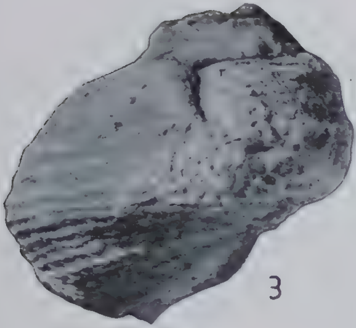
TYPES

HOLOTYPE: Specimen TM 2055, GS 6661, figs. 16, 17, 23.

PARATYPES: Specimen TM 2056, GS 6075, TM 2053 and two at the Geology Department, University of Otago, CG 6 and CG 7, GS 5519.

TABLE 2. Dimensions (in millimetres) of specimens of *Mellarium nodulosum* kept at University of Otago

Specimen	Width W	Height H	Height Body Whorl B	H/ /W	B/ /W	Width Aperture	Pleural Angle
GS 6661 (holotype)							
TM 2055	27	? 23·5	19·5	0·87	0·83		120
GS 5519							
CG 6	20	23	17·5	1·15	0·74	9	95
CG 7	22	? 22	16·5	1·0	0·75	10	90
TM 2053	25	22	12·5	0·88	0·57	11	110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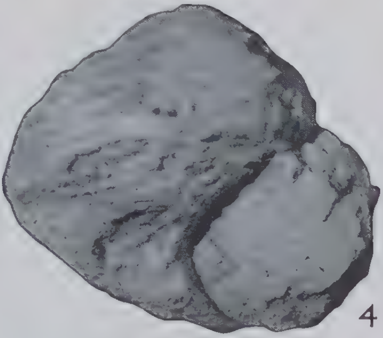
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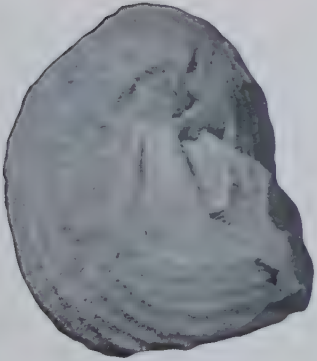
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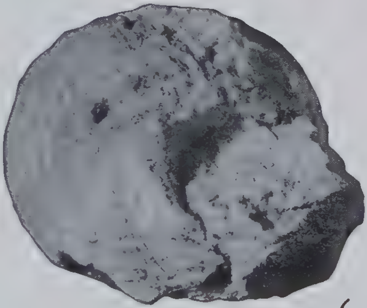
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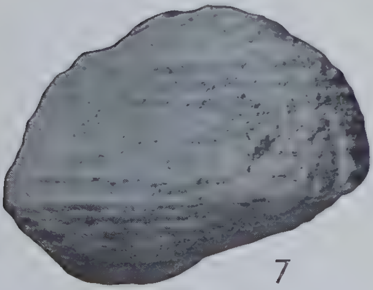
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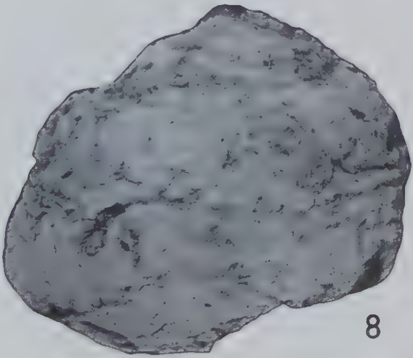
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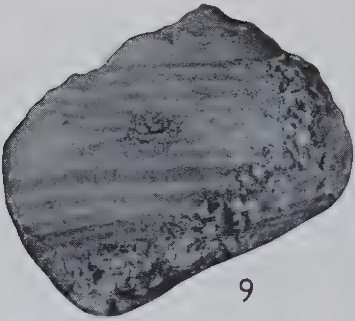
Mellarium mutchi n. sp.

(All specimens from GS 6066, × 1·5 approx.)

FIGS. 3, 4, 6—Lateral, apertural and basal view of the holotype TM 2040.

FIGS. 5, 7—Basal and lateral views of a squashed specimen TM 2045, with a very large body whorl.

FIG. 8—Lateral view of polyvinyl chloride mould of specimen TM 2051, with a very large body whorl, like that of TM 2045.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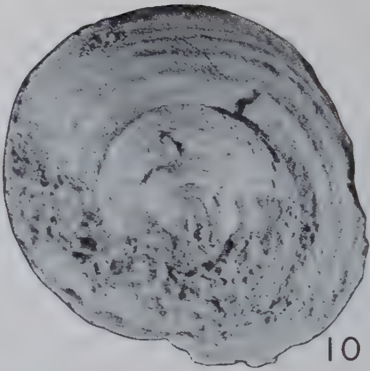
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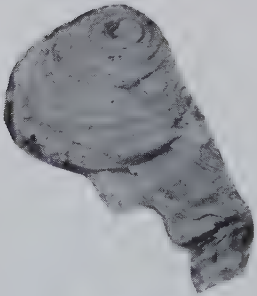
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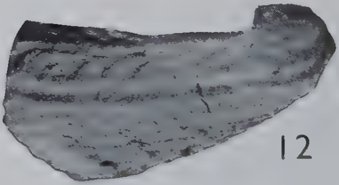
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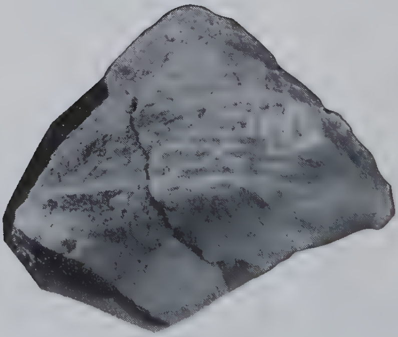
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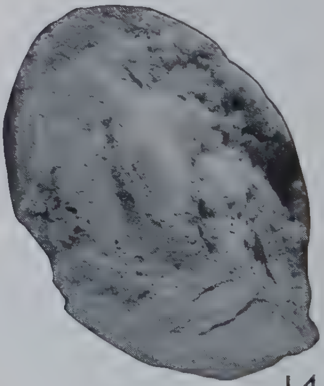
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Mellarium mutchi n. sp.

(G.S. 6066, × 1·5 approx.)

FIGS. 9, 10—Lateral and top view of low-spired paratype, TM 2048. Polyvinyl chloride mould.

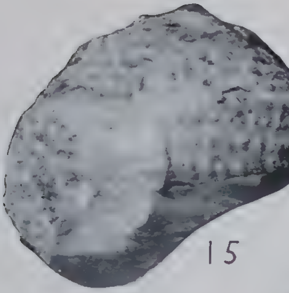
FIG. 11—Oblique view of incomplete specimen showing ornament and selenizone. Polyvinyl chloride mould.

FIG. 12—Lateral view of incomplete specimen, showing selenizone and underlying part of the body whorl. Polyvinyl chloride mould.

Mellarium cf. *mutchi*

(GS 4514, × 1·5 approx.)

FIGS. 13, 14—Lateral and basal view of distorted specimen TM 2054. Fig. 13 shows the aperture in side view, with rock matrix attached.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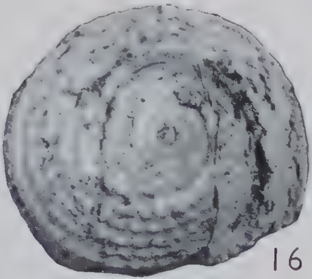
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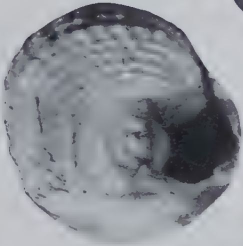
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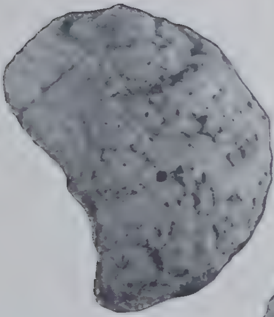
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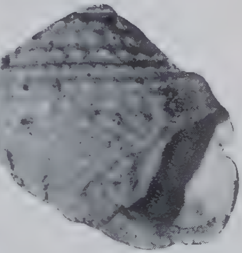
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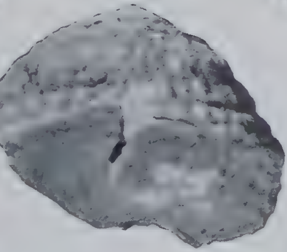
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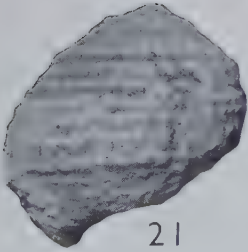
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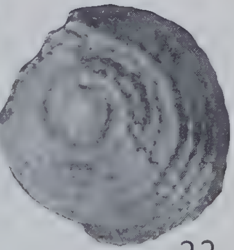
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Mellarium nodulosum n. sp.
(All specimens, × 1·5 approx.)

FIGS. 15, 16, 18—Lateral, top, and oblique views of holotype TM 2055, GS 6661. Polyvinyl chloride mould.

FIGS. 17, 19—Basal and apertural views of TM 2056, GS 6095. Polyvinyl chloride mould.

FIG. 20. Apertural view of paratype TM 2053, GS 5519.

FIGS. 21, 22—Lateral and top views of paratype CG 8 (University of Otago). G.S. 5519.

TABLE 3—Occurrences of *Mellarium* in New Zealand

Collection Number	Description of Locality	Age
Mellarium mutchi		
GS 3587 S 159/ 489	Upper Beaumont Scarp, 64 chains at 255° from Trig. D. N.Z. map Series, Provisional Sheet S 159, 1953 ed., grid. ref. 029668.	Anisian
GS 3593 S 159/ 488	Lower Beaumont Scarp No. 1. Left bank trib. of North Branch of Beaumont Stream, rising at Beaumont Trig. Grid ref. 024669.	Anisian
GS 5199 S 159/ 443	Isolated small hill, 30 chains west of Spot Height, 1,280 ft, North Branch, Beaumont Stream. Grid ref. 022654.	Anisian
GS 6326 S 159/ 670	Scattered outcrops on the top crest of the ridge at western edge of beech forest, midway between Barrett's Hut and Coral Bluff, on a saddle between two eastern branches of Letham Burn, Grid ref. 991790.	Lower Anisian
GS 6066 S 159/ 620	Ten chains east, towards Barrett's hut, from the preceding locality, in small outcrops and boulders on the south side of a grassy ridge, S.D. Grid ref. 996791.	Lower Anisian
S 179/ 655	Wave-cut platform about 30 yd S of Kaka Point sandstone promontory. N.Z. map Series, Provisional sheet S 179, 1944 ed., grid ref. 567508.	Anisian
Mellarium aff. mutchi		
GS 4514 S 131/ 501	South end of Countess Range 135 chains at 158° from Winton Peak, below Spot Ridge "5,210 ft" on map. N.Z. map Series, Provisional sheet S 131, grid ref. 023543.	? Triassic
Mellarium nodulosum		
GS 5519 S 159/ 542	South end of second prominent high bluff north of first east tributary of North Etal Creek. 240 chains at 352° from Trig. E + Etal. Wairakei S.D. N.Z. map Series, Provisional sheet S 159, 1945 ed., grid ref. 046801.	Upper Anisian
GS 6075 S 159/ 627	Second east branch of North Etal Creek in prominent knob on south leading ridge to third and southern height of Etal. Grid ref. 044743.	Upper Anisian
GS 6661 S 159/ 754	Three chains east of fence and 25 chains south along fence from corner nearest to the east branch of Beaumont Stream and old sawmill. Grid ref. 042639.	Upper Anisian
Mellarium aff. *nodulosum* (from 100 ft below *M. mutchi* at S 179/655).		
S 179/ 657	Nearshore edge of wave-cut platform south of Kaka Point promontory, about 60 yards south of northernmost flinty siltstones. N.Z. map Series, Provisional sheet S 179, 1944 ed., grid ref. 567508.	?

DESCRIPTION. The shell is of moderate size, comprising five globular whorls ornamented with nodules. Sutures are abutting, and the base concave and cryptomphalous. The early whorls are obscure, but appear to be more globular than the third and succeeding whorls, and have a low spiral angle measuring 90° to 100°. The last three whorls embrace each preceding whorl more, up to the upper edge of the selenizone, except for the last half turn of the body whorl, which lies at the lower edge of the selenizone. The body whorl is sub-globular in transverse profile, with three angulations, the uppermost at the outer rim of a narrow but prominent shelf, a second weak angulation at the selenizone, which lies just above mid-height on the whorl, and a third prominent angulation immediately above the base. The periphery lies at the basal angulation near the aperture, and between the basal angulation and selenizone away from the aperture. Between the sutural shelf and the selenizone the whorl profile is concave; below the selenizone the profile is gently convex.

Parietal callus is not developed. The aperture is prosocline, sloping as a whole at only about 30° from the suture. The upper apertural margin slopes at 60° to 70° from the suture: below the selenizone the margin curves forward at a very low angle from the selenizone, to the immediately underlying row of nodules, and then turns abruptly to sweep with slight convexity towards the aperture, and almost vertically across the lower outer profile. The lower aperture passes prosoclinally at about 55° from the lower periphery and is gently sinuous in outline, being convex towards the aperture near the periphery and near the columella, and concave over the median part of the base.

Judging from strong growth-lines on two specimens, the slit is about one-fifth of the circumference of the shell, measured from the anterior upper margin. The selenizone is gently convex and ornamented with sickle-shaped nodules. It is sharply defined by a narrow groove on each side.

The ornament comprises nodules in spiral rows. Four rows occur on the body whorl above the selenizone, three rows occur between the selenizone and the lower periphery, and another seven narrower and more closely spaced rows lie over the base. Each nodule is low and boss-like, and generally slightly elongated towards the aperture. Generally the nodules on the upper whorl profile are best defined; a low ridge may connect the rows below the selenizone, and on the base each row of nodules is connected by a low costa. The nodules are also aligned between growth-lines, particularly on the base, where the growth-lines form deep slits between the radial rows of nodules. Nodules are well developed also on the penultimate and next whorl, but nodular costae are present on the early whorls.

The shell is 0·5 mm thick.

VARIATION. Insufficient specimens are available to discover the variability of the species. The difference in size and shape recorded in the measurements are due at least in part to crushing.

One hundred feet below *M. mutchi* at Kaka Point is a specimen CG 9 with a nodular selenizone as in *M. nodulosum*, but with costae as in *M. mutchi*. Thus in some respects, it is intermediate between the two species, but its position below *M. mutchi* discourages a picture of any simple development of *M. mutchi* into *M. nodulosum.*

RESEMBLANCES. This species is distinguished from *M. mutchi* by the ornament of nodules. In addition, it has a lower spire, and the selenizone is overlapped by succeeding whorls. The sutural shelf and basal angulation are more prominent, the profile more concave above the selenizone, and the upper and lower margins of the outer aperture less rounded towards the aperture.

Of upper Paleozoic species of *Shansiella*, the closest is *S. carbonaria*, which has a somewhat similar angular whorl profile, but a stronger basal angulation and more rounded sutural shoulder.

AGE. This species occurs in Anisian beds mapped as upper Etalian by Mr A. R. Mutch (pers. comm.).

ACKNOWLEDGMENTS

The writer is grateful to Dr E. L. Yochelson, United States Geological Survey, Washington, D.C., for discussing the relationship of *Mellarium* to *Shansiella*, and to Dr C. Waterson, Royal Scottish Museum, Edinburgh, Scotland, who kindly made facilities available for examining Scottish specimens of *Shansiella* and later sent latex moulds of the types of the Scottish species. Literature not available in New Zealand was seen during the tenure of an 1851 Royal Exhibition Scholarship at Cambridge, England.

The writer also wishes to thank Mr J. D. Campbell, University of Otago, for stratigraphical information, and for the loan of specimens, and Messrs A. R. Mutch and G. W. Grindley, New Zealand Geological Survey, for stratigraphical information.

Photographs are by Mr S. N. Beatus, and the graphs were drafted with the help of Mr C. T. Webb, New Zealand Geological Survey.

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NOTE ON THE GORE-BALFOUR SERIES BOUNDARY

By J. D. CAMPBELL and D. S. COOMBS, University of Otago, C. A. FLEMING, D. KEAR, I. C. MCKELLAR, A. R. MUTCH, I. G. SPEDEN, J. B. WATERHOUSE, and B. L. WOOD, New Zealand Geological Survey, and J. MARWICK.

(Received for publication, 28 September 1959)

Series in the Triassic were introduced by Marwick (1951) "largely [for] . . . mapping requirements", a grouping of several stages being a more mappable unit than a single stage, especially on small scale maps. The decision of the New Zealand Geological Survey to use series as the units in the 4-mile mapping project bears out the usefulness of these units.

It is probably true to say that in general, boundaries between New Zealand series have no greater significance than boundaries between any two stages, and this is consistent with Marwick's comment quoted above. Where possible, however, series boundaries might be used to emphasise distinct faunal events even though in general they are arbitrarily placed.

Marwick placed the boundary between the two Triassic series, Gore and Balfour, at the Oretian-Otamitan stage boundary. This gives a division in thickness in several sections, Gore Series first, as follows: 25,650, 9,850; 8,220, 3,280; 15,850, 6,650; 9,900, 2,170 (all figures in feet based on Coombs, 1950; Mutch, 1957). It is obvious that Gore and Balfour series as constituted at present are mappable units. It should be noted here that Gore Series, given no lower limit by Marwick (1951), and exended down only to the base of the Etalian by him in 1953, has more recently been used for all pre-Otamitan Triassic (Mutch, 1957).

Gore Series as now used accommodates Oretian, Kaihikuan and Etalian stages as well as the pre-Etalian Malakoff Hill beds (Kummel, 1959) and pre-Kaihikuan North Range beds (Coombs, 1950) and Wairuna Peak beds (Wood, 1956). The consistently great thicknesses for the Gore Series together with its inclusion of rocks of Lower Triassic age will almost certainly lead to its subdivision [compare Mutch's (1957) use of "Lower Gore Series"] with a new series to include beds with Lower Triassic fossils. A revision of the content of Gore Series may thus be expected in the future.

With the 4-mile mapping project in hand it is timely to review the series even though the lower limit of the Gore Series cannot be fixed at present. The Oretian-Otamitan boundary subdivides the Triassic into mappable units, but its choice as series boundary is unhappy on several grounds.

The entry of *Mytilus problematicus* at which the lower boundary of the Otamitan is drawn is a convenient and readily recognised datum in all but a few sections. The entry of *Halobia* at the base of the Oretian is, however, a much more significant event faunally because of this genus's wide distribution and hence its use in inter-regional correlation. *M. problematicus*, on the other hand, appears to be restricted to the Maorian Province.

In the New Zealand sequence the best known sections in Triassic rocks are those across the northern limb of the Southland Syncline. The most distinctive faunal break in succession after succession there comes at the Kaihikuan-

N.Z. J. Geol. Geophys. 3 : 284–286.

Oretian boundary. It is doubtful if any molluscan or brachiopod species ranges across the boundary. Kaihikuan and Oretian assemblages can be readily distinguished even when *Daonella* and *Halobia*, the two key fossils, are not present. Oretian and lower Otamitan faunas, on the other hand, have much in common and undifferentiated Oretian-Otamitan has been mapped (Wood, 1956, p. 62) where *M. problematicus* could not be found.

In view of the difficulty of differentiating Oretian and Otamitan faunas, it would be more satisfactory to have these stages included in one series. For this reason and in order to emphasise the faunal break at the base of the Oretian it is proposed that the Gore-Balfour series boundary be placed between the Kaihikuan and Oretian stages.

Revised Subdivision of Triassic System in New Zealand

Map Symbol	Stage	Series
Bo	Otapirian	Balfour
Bw	Warepan	
Bm	Otamitan	
Br	Oretian	
Gk	Kaihikuan	Gore
Ge	Etalian	
	Pre-Etalian Triassic	

It should be noted that the Kaihikuan-Oretian boundary correlates broadly with the Ladinian-Carnian boundary of the European Alpine Triassic so that Balfour series is equivalent in a general way to Spath's (1934) Neo-Trias but that this near correspondence of New Zealand and European units is an indirect result of the change and not a major reason for it.

TRIASSIC SERIES AND THE ALPINE FACIES

The classification of Alpine Facies greywackes of Canterbury and elsewhere in Gore and Balfour Series as originally defined poses a special problem. In some areas they could formerly be assigned on fossil evidence to one series or the other, e.g., Gore for rocks with Kaihikuan fossils, Balfour for rocks with *Monotis*, but beds with *Halobia* could not be distinguished as either Oretian (Gore) or Otamitan (Balfour). With the amended grouping of stages the Balfour Series will accommodate all beds with *Halobia*. For the important *Terebellina* beds, exact placing is still uncertain. On several lines of evidence, although none incontrovertible, a pre-Warepan and post-Oretian age is suggested (Campbell, MS) for these beds. The change in boundary allows accommodation of them in the Balfour Series.

MAP SYMBOL

The symbol Br is proposed for the Oretian Stage.

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STRUCTURAL FEATURES OF THE ALPINE SCHISTS OF THE FRANZ JOSEF – FOX GLACIER REGION

By Bernard M. Gunn, N.Z. Geological Survey, Christchurch*

(Received for publication, 13 October 1959)

Summary

Fold axes in the steeply dipping greywackes and schists to the west of the Alpine divide may be located by determinations of the superposition of beds made from observations of: (1) graded bedding; (2) intraformational structures; (3) contemporaneous deformation structures; (4) crossbedding; (5) slip folds; (6) drag folds and minor folds; (7) axial plane cleavages, joints and quartz veins; and (8) feather joints. Both joints and fault patterns have been demonstrated to be statistically related to the fold axes by means of equal area projections. The fold axes trend between 40° and 50° and locally diverge southward from the Alpine fault whose strike is 58°. Petrofabric study of calcite optic axes from marble shows a preferral orientation nearly normal to the Alpine fault. The direction of maximum principal stress during deformation inferred from fold axes and joints is about 310°, while that inferred from the attitude of the Alpine Fault and the calcite orientation is about 327°, but it is not thought that the difference is significant.

Introduction

As a result of recent work the structural plan of the central part of the New Zealand Southern Alps is beginning to emerge and a summary of results up to the year 1955 has already been published. (Lillie, Gunn and Robinson, 1957.)

To the east of the Alpine divide, the exposures of closely folded Alpine greywackes are ample for structural details to be elucidated. To the west, however, a heavy forest cover is virtually continuous and exposures are limited. In two valleys only, those of the Franz Josef and Fox Glaciers, are there continuous, freshly ice-scoured rock walls showing the detail needed to follow the pattern of folding. In this paper the structure of this region is described in detail, as even here the folds are nearly isoclinal and the fold axes cannot be located without an understanding of several common structural indicators.

The metamorphic grade of these rocks changes progressively from the muscovite-chlorite subfacies (greywacke and argillite) at the Alpine Divide to schists of the amphibolite facies eleven miles to the north-west where the Alpine schists are cut off by the Alpine Fault. As might be expected, the criteria useful for structural interpretations vary with the degree of metamorphism of the schists.

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N.Z. J. Geol. Geophys. 3 : 287–308.

FOLD STRUCTURES

Criteria Used to Determine Superposition of Beds (fig. 1)

(a) *Graded Bedding*

Graded beds are common in the greywacke zone, and may be recognised in the chlorite zone schists and even in the early biotite schists. Usually the graded beds occur in groups up to 20 ft thick, with each bed seldom more than 4 in. thick. The grain size may grade from fine sandstone upwards into dense, black mudstone within the confines of a single bed or the variation may be so slight as to show only as a change in colour from light to darker grey.

(b) *Intraformational Structures*

Fragments of argillaceous rock are sometimes seen embedded in greywacke. The fragments are often bent and are interpreted as having been caught up by a slump or turbidity current flow of sand in a geosynclinal environment. When the sandstone bed is several feet thick and the argillaceous fragments are embedded within inches of a contact with an argillaceous bed, the sandstone is taken as being the younger.

(c) *Contemporaneous Deformation Structures*

Slumping of soft clays has in some places produced asymmetrical interbedded folded structures and dislocations which reveal which surface was the upper. A photograph of such a slump structure has already been published (Lillie, Gunn and Robinson, 1957).

(d) *Cross Bedding*

In siltstone beds near the main divide occasional examples of cross bedding occur in laminae, less than 0·1 in. thick. The topset beds are frequently truncated and so the structures are concave upwards. Though rare, such structures have formed a useful check on graded-bedding determinations. Scour pits are also occasionally present.

(e) *Slip Folds*

These occur in a restricted zone of early biotite schists and also near fold axes in the greywackes. Slip folds can only occur when the dip of the slip planes deviates from the dip of the bedding. Slip of the core of a fold outward along axial plane cleavages in a late stage of folding has produced steplike displacements of the beds. If the slips are close enough together, the slip fold simulates a drag fold in outline. The presence of drag folded axial-plane quartz veins show that such outward slip has taken place. The intersection of the slip planes with the bedding imparts a lineation which may plunge steeply if the strikes of the bedding and of the slip planes are not parallel.

(f) *Drag Folds and Minor Folds*

These are most conspicuous in biotite and higher grade schists. Drag folds are defined (Billings, 1955, p. 45) as being the internal deformation produced in an incompetent bed, subject to shearing between two competent beds. By Pumpelly's Law the attitude of these minor folds is congruent with the attitude of the major folds. True drag folds are found in the low-grade

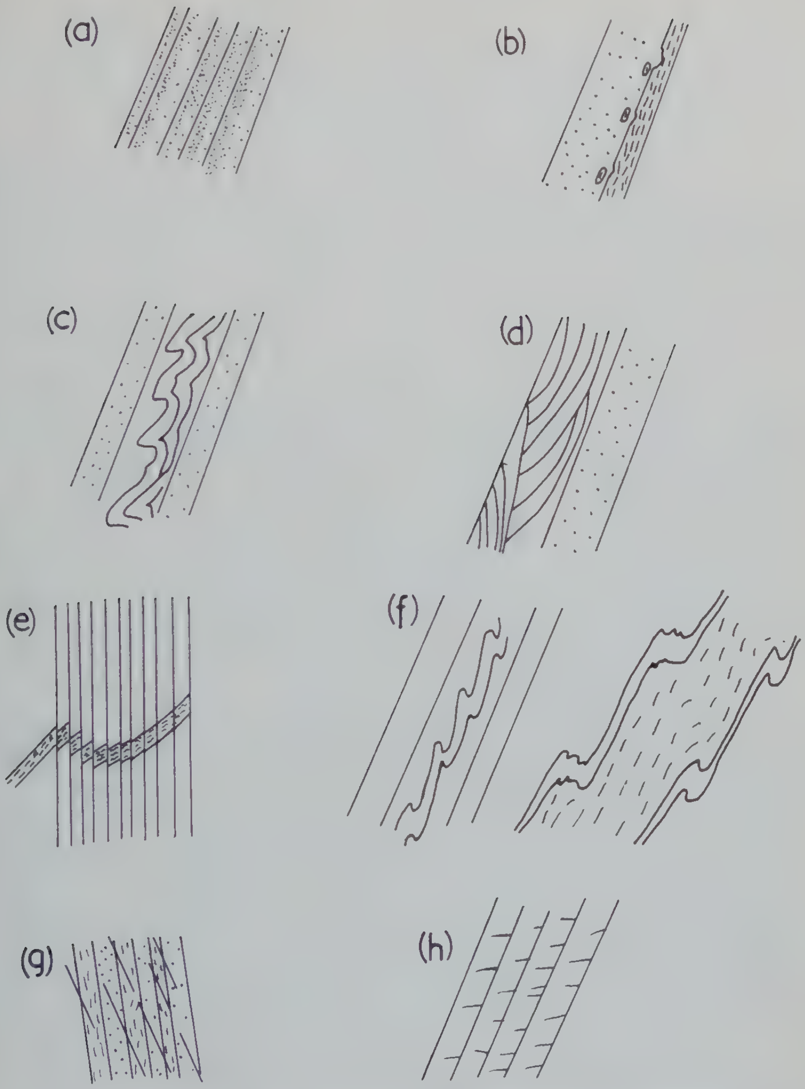

FIG. 1—Diagrams of: (a) Graded bedding; (b) Intraformational structures; (c) Contemporaneous deformation structures; (d) Cross bedding; (e) Slip folds; (f) Drag folds and asymmetric minor folds; (g) Axial plane quartz veins; (h) Feather joints.

In each case the anticline axis is to the right.

schists where argillaceous beds are dragged between competent indurated sandstone beds, but in the higher grade schists it is the relatively competent quartz segregations that are folded on a small scale while the incompetent micaceous segregations show flowage into the apices of the folds. These are minor folds rather than true drag folds but they are formed in the same way and have the same significance. The amplitude of the minor folds within about 300 ft of a major fold axis may exceed 1 ft, but this decreases both towards the fold axis and toward the limb of the fold.

(g) *Axial Plane Cleavage, Joints, and Quartz Veins*

In the absence of drag folds, a cleavage which is often rendered more prominent by large parallel flakes of biotite, is often found parallel to the axial plane of the nearest major fold. The bedding then crosses the cleavage upwards in the direction of the anticlinal fold. Axial plane joints have a similar significance.

In a broad zone near the terminus of the Franz Josef Glacier, parallel lenses of quartz oblique to the bedding are prominent. These lenses are usually between $\frac{1}{4}$ and $\frac{1}{2}$ in. thick, 6–18 in. high, and up to 4 ft long, and may exceed 1 per square foot of surface in density. These quartz veins were found to be parallel to the axial plane of the nearest fold and to have the same significance as the axial-plane cleavages elsewhere.

(h) *Feather Joints*

Shearing stresses between beds near the axes of folds in the greywackes and low-grade schists has resulted in the formation of tension joints inclined at a high angle to the bedding. The acute angles between joint and bedding points in the direction of relative displacement.

Structure of the Upper Franz Josef and Fox Valleys

The alpine greywackes west of the central axis of the Southern Alps have a uniformly steep westerly dip with a strike varying between 015° and 040°. Graded bedding determinations, all indicating the tops of the beds to be to the west, were made at Mackay Rocks, on the Lower Linden Ridge, at Teichelmans' Corner, on the Almer Ridge above Almer Hut, on Pioneer Ridge above Pioneer Hut and also at two points west of the hut, at the foot of the northern ridge of Big Mac, on the extreme eastern end of Chancellor Ridge and at the eastern end of Middle Deck, Chancellor Ridge.

Intraformational structures also indicating tops to the west were found at Pioneer Ridge, Westhoe Pass and McKay Rocks. Slip folds plunging 32°S and indicating a syncline to the west were found at the eastern end of Middle Deck, and a vertical axial plane cleavage intersecting the westward dipping bedding on the Almer Ridge, Pioneer Ridge and on the north ridge of Big Mac has a similar significance. The intersection of bedding and cleavage on Almer Ridge plunges southward at 55°. The rocks of the Fritz Range form an almost unbroken sequence from east to west and show no folding east of Mount Roon. A minimum thickness of some 20,000 ft of strata is therefore involved.

The Almer - Middle Deck Syncline

Eastward from Cape Defiance, Franz Josef Glacier, the limbs of well defined drag folds in the biotite schists dip 75°E. The drag folds indicate a syncline to the east. Near the foot of the main icefall, rather poorly defined graded bedding shows the beds to be rightway up with tops to the east. The major synclinal axis lies at the foot of the icefall and has reverse folds up to 15 ft in amplitude in the apex. The western limb dips eastward at about 60° and the eastern limb is nearly vertical but dips less steeply farther to the east. At the level of the glacier, the reverse folds extend over a distance of more than 200 yd before the bedding finally turns upward once more. The plunge of the secondary and tertiary folds varies between 25° and 40° south. When the poles of the bedding from each side of the fold are plotted on a stereographic projection an irregular curve is produced about a plunge of 35°S. The plunge is therefore thought to be real and not merely due to a divergence between bedding and axial plane schistosity. This syncline will be referred to as the Almer Syncline although the axis passes rather more than a mile west of the Almer Glacier.

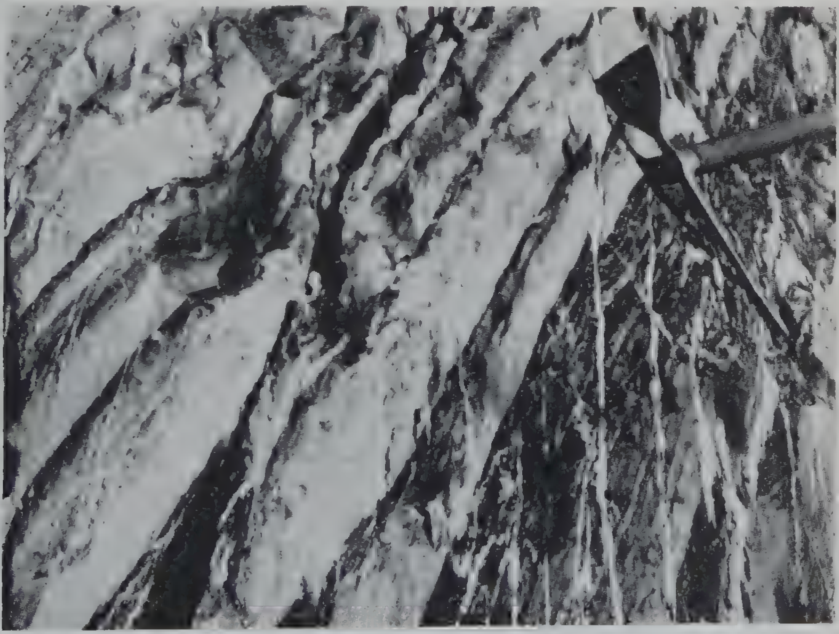

FIG. 2—Photograph of early biotite schists, Middle Deck, Fox Glacier showing eastward dipping bedding, vertical axial-plane quartz veins and cleavages, slip folds and irregular drag folds and graded bedding all showing tops to the left.

A similar syncline is found at Middle Deck, Fox Glacier. Its presence is indicated on the eastern limb by the westward dipping bedding, graded bedding showing tops to the west, and by slip folds and feather joints showing upward displacement to the west. On the western limb the presence of a synclinal fold is indicated by drag folds grading into slip folds which show upwards displacement to the east, graded bedding showing tops to the east, and eastward dipping strata. The bedding at Point A on the western limb is sheared and irregular, but has a general dip of 58°E. Quartz veins $\frac{1}{4}$ in. thick apparently parallel to the axial plane of the fold, cross the bedding and stand vertically. The apex of the syncline can be seen on the ice-scoured rock 600 ft above the Fox Glacier and reverse folds occur for about 50 yd on each side of the axis. As in the Almer syncline, the axial plane cleavage dips vertically or 80°E and the intersection of cleavage and bedding plunges south at 10°. Fifty yards to the east of the syncline the dip of the beds is 50°W and on the summit of Chancellor Dome it is 70°E. The dip reverts gradually to the general westerly direction east of the Chancellor Dome and as at Franz Josef this regional change in dip is not closely associated with folding.

The Almer syncline lies in a direction of 030° from the Mobile Deck syncline and as this is almost exactly parallel to the strike of the bedding, there can be little doubt that only one continuous fold is present.

Westward of this syncline, the folds of the Fox and Franz Josef Valleys are less similar and will be described separately.

Folds of the Lower Franz Josef Valley (fig. 3)

The Cape Defiance Flexure

At Cape Defiance and also on the opposite bank of the glacier a belt of disharmonic folds each 5–20 ft in amplitude is found. The bedding planes of individual folds are not parallel due to extensive thickening of the micaceous layers and to a less extent, of the quartz segregations, towards the crests of the folds. Eastward at Harpers Creek the minor folds become asymmetric with longer eastern limbs so that the average dip is about 50°E and this dip becomes steeper towards the Almer Syncline. Westwards towards Arthurs Gateway the minor folds become asymmetric in a similar way until the bedding dips 70°E with regular drag folds of 1 in. amplitude and 5 in. wavelength. This flexure between Harpers Creek and Arthurs Gateway is about half a mile in width and may form a primary anticline and syncline at greater depth or may merely represent a large scale drag fold.

The Roberts Point Anticline

A quarter of a mile west of the end of the Roberts Point track and 1½ miles north-west of the Almer syncline the bedding inverts in a tight anticlinal fold. On the eastern limb the bedding dips 65°, strike 135°, and axial-plane quartz veins dip 70°E, strike 005°. Drag folds reach a maximum amplitude of 1–6 in. 150 yd east of the fold and consistently indicate an anticline to the west. The plunge is sometimes as much as 60°S. The western limb is overturned with a dip of 70°W 200 ft from the axis but with a dip of 70°–80°E at Rope Creek and steep easterly dips are found for a mile and a half to the west. On this overturned limb drag folds are rare but axial-plane quartz veins dipping 45°E are plentiful. At Dolly Creek the bedding once more is right way up with a dip of 75°NW, strike 045° and drag folds have again appeared indicating a syncline to the west.

The Callery Gorge Syncline

The synclinal axis passes under the foot bridge at the Callery Gorge. The bedding is shown to be parallel to the original lamination by green hornblende

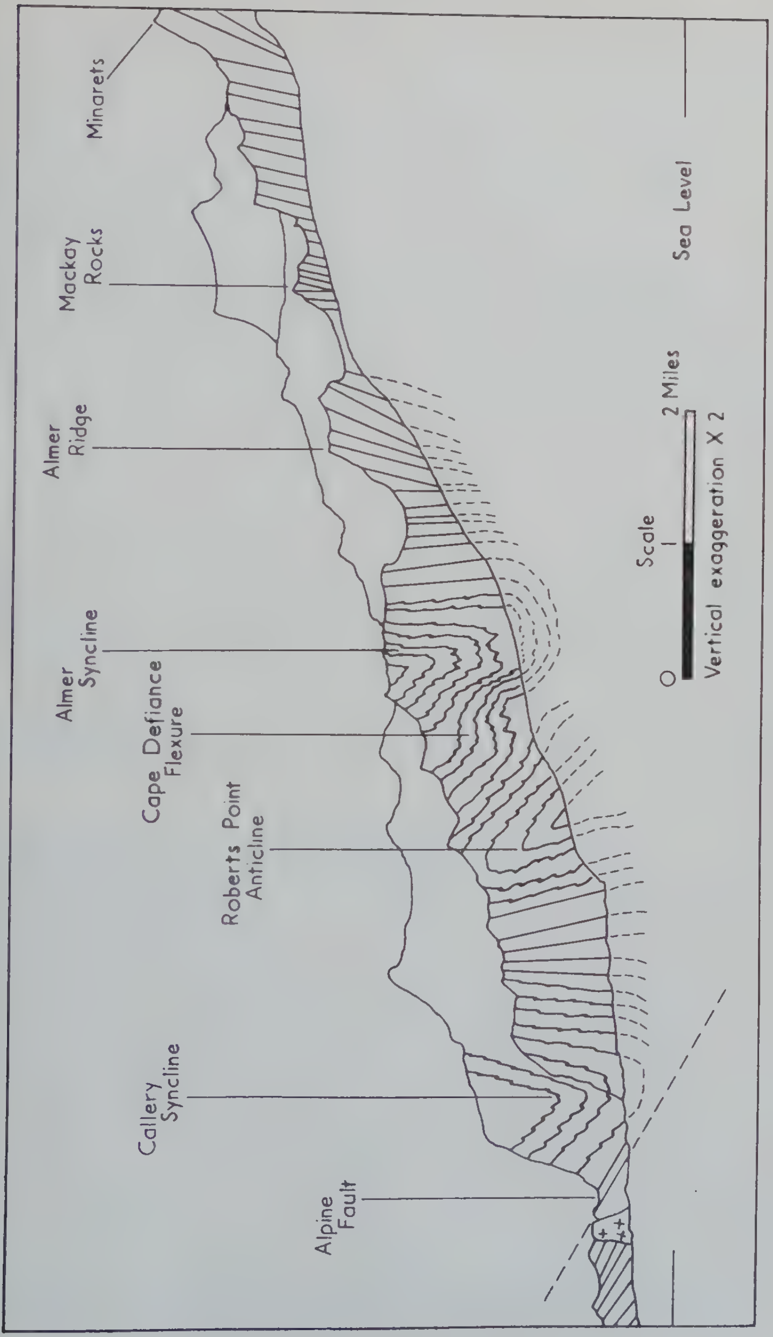

FIG. 3—Folds of the Franz Josef Valley projected normal to fold axes.

garnet schist interbedded with quartzo-feldspathic biotite schist. The beds are inverted over a general curve of about 300 ft diameter but this contains reverse folds of an amplitude of about two feet. The axial plane cleavage dips at 75° to 78°E and the plunge of the secondary folds is 14° to 18° at 220°. About 200 ft west of the axis the bedding lamination dips east at 64°.

In the Tartare Gorge near the hydroelectric intake, garnet schists are strongly dragfolded and indicate a syncline to the east. The wavelength of the folds varies between 6 in. and 2 ft with multiple apices with a total amplitude of about a quarter of the wavelength. The limbs of the dragfolds dip eastward at 58° to 74° and the Callery syncline probably passes a short distance east of the intake. The gorge is, however, completely inaccessible.

In the vicinity of the Alpine Fault strong faulting and shattering of the rock prevents further diagnoses of folding. The lamination of the schists immediately east of the Alpine Fault dips south-eastward 38°, strike 048°. These rocks are massive but may have been disoriented by the surrounding faults.

Folds of the Lower Fox Glacier Valley

The fold sequence here is similar to that previously described but the folds are on the whole more open and low angle dips are more common. The axial plane cleavage or schistosity varies by less than 10° from vertical throughout the entire sequence.

The Paschendale Ridge Anticline

West of the Middle Deck syncline, the bedding continues to dip eastward at 70° or less, intersecting the axial plane quartz veins and joints which dip 85° to 90°E.

On the south bank, 200 yd east of Paschendale Bluff and two miles west of the Middle Deck syncline the bedding abruptly curves over and becomes horizontal but 2,000 ft above the glacier the curve is much more regular and gentle, describing an arc centred somewhat below glacier level. Along the crest of the Fox Range on the east side of Sam Peak a continuation of this flexure was found to occur about a much smaller radius of curvature and with a similar plunge of 30°S.

From Paschendale Bluff westward to Cone Rock, a distance of about 1 mile 50 ch across the strike, the sequence can best be followed along the north bank of the glacier. The bedding remains sub-horizontal but is strongly crumpled, commonly with tertiary folds of 6 in. to 1 ft amplitude and up to double this wavelength. Quartz segregations in the apices of folds may be 5 in. thick but completely sheared out on the limbs. Each limb is harmonically drag folded, and, as at Franz Josef, few folds are repeated vertically for more than a few feet without their shape being completely altered or even becoming inverted. The mean direction of these tertiary folds curves up and down forming open secondary folds of 6–10 ft amplitude and these again are part of larger structures which, however, do not appear to have a greater amplitude than a few hundred feet. On the south bank prominent interbedded green-schist lenses lie almost horizontal. One such lens extends for almost a mile east of Boyd Creek and varies in thickness up to a maximum af nearly 200 ft. It lies up to 1,000 ft above glacier level and at this elevation the schist is less strongly crumpled and the bedding is regular and nearly horizontal.

Halfway between Cone Rock and Fork Creek, the bedding is shown on the Cone Rock fault-line scarp to be crumpled but to have a general dip of 60°W with a plunge a few degrees on each side of horizontal.

The Cone Rock Syncline

At the eastern end of Cone Rock the bedding becomes sub-horizontal and near the old glacier gallery it sweeps upward to assume an attitude dipping 74–80° E, strike 043° which is maintained until the western end of Cone Rock is reached. Here another minor flexure occurs, the bedding becomes sub-horizontal for a distance of 40 yd and then reverts to an easterly dip of 67°. These rocks near the terminus of the Fox Glacier are similar to those of the lower Franz Josef and have a single prominent lamination which is parallel to the macroscopic single schistosity but with rare sheared-out dragfolds indicating a syncline to the east. Axial plane quartz veins dip about 80° eastward and these are also rare.

On the north wall of the glacial trough opposite Cone Rock the plunge of the dragfolds is 22° to 35° to the north in contrast to the plunge of 20°S at the western end of Cone Rock. Rock exposures become increasingly rare towards the western escarpment. In Rocky Creek the rock is much shattered by faulting and the rock sequence is discontinuous but near the Alpine fault a general easterly dip of 34° to 55°E prevails.

In the Waikukupa River east of the Alpine fault, the lamination dips to 60°E, strike 035° but faulting is intense and this may well be not parallel to the original bedding.

Regional Trends of the Primary Folds

If the folds of the Franz Josef valley are joined on the map to those of the Fox, the directions of their axes are nearly parallel to that of the general bedding strike. It seems likely that these folds are continuous for this distance of 12 miles at least.

The strikes of the axial planes of the folds are as follows:

Almer-Middle Deck syncline	036°
Roberts Point – Paschendale anticline	046°
Callery – Cone Rock syncline	046°

Plunge of Fold Axes

Both Morgan (1908, p. 78) and Bell and Fraser (1906, p. 42) comment on the plunge in the fold axes and attribute this plunge to the effect of an earlier folding movement at right angles to the present mountain axis. As stated above the plunge may exceed 60° and theoretically it seems questionable whether a sequence of rocks could be folded so strongly and then be folded in a transverse direction and retain the regularity of structure that has been suggested. A curious feature is that although the limbs of folds diverge in dip by at least 25° in each case seen and therefore are not isoclinal, the general strike of the beds on each limb remains parallel except for a distance of a few hundred yards on either side of the axis.

Unfortunately, rock exposures are very limited along the strike and it is not known whether the plunges of the lineations and small scale folds are parallel to the large-scale features. This is one of the biggest problems yet to be solved.

Fold Structures of the Greenland Group Greywackes

The Greenland Group greywackes lying to the west of the Alpine Fault are exposed only in restricted, widely-separated areas and no complete picture of their mode of folding could be built up. In general, the dip of the Greenland beds is less than 45° and so they have been assumed to be right way up. No graded bedding, drag folding, cross bedding or other structural indicators were found and they appear to be completely unfossiliferous.

An equal area projection of 20 bedding poles from three localities suggests folding about an axis striking 10° west of north. Near Waiho along Gibbs Road and Gibbs Creek the beds dip 45° eastward, in Pratt Creek, a tributary of the lower Tartare, the beds are nearly horizontal, while in McDonalds Creek, 6 miles from Gibbs Road, a maximum westerly dip of 43° is found. The structure is that of an open syncline striking 350° and plunging about 10° to the north. The strike of individual beds varies by 15° on each side of this mean position and with so few outcrops, this is at best an approximation. Other folds may be present between the limits of the visible outcrops and hidden beneath the coastal moraine and gravels.

JOINT SYSTEMS

Mackay Rocks Joint Pattern

Equal area projections of joints recorded in the field from selected localities show the joints to have a consistent pattern which can be related to the major fold structures

The projection of joints from the greywackes of Mackay Rocks, Franz Josef Glacier is shown in fig. 4 and is contoured about concentrations of 5%, 10%, and 20% per unit area. As many different orientations as possible were recorded in the field and so the actual numerical preponderance of the concentrations is therefore even higher than the diagram indicates. Two strong maxima shown include 85% of all points.

Poles of the bedding show a close grouping about a mean strike and dip of 031°/75°W and a small group of bedding joints is superimposed on this. The strongest concentration of 25% lies near the perimeter; the joints are normal to the bedding and stand vertically. As such joints are commonly perpendicular to the axes of the folds, the tectonic axis (*b*) is probably almost horizontal (Billings, 1955, p. 117). A plunge northward of 10° is suggested but the number of joints recorded from this locality is too small for this figure to be significant.

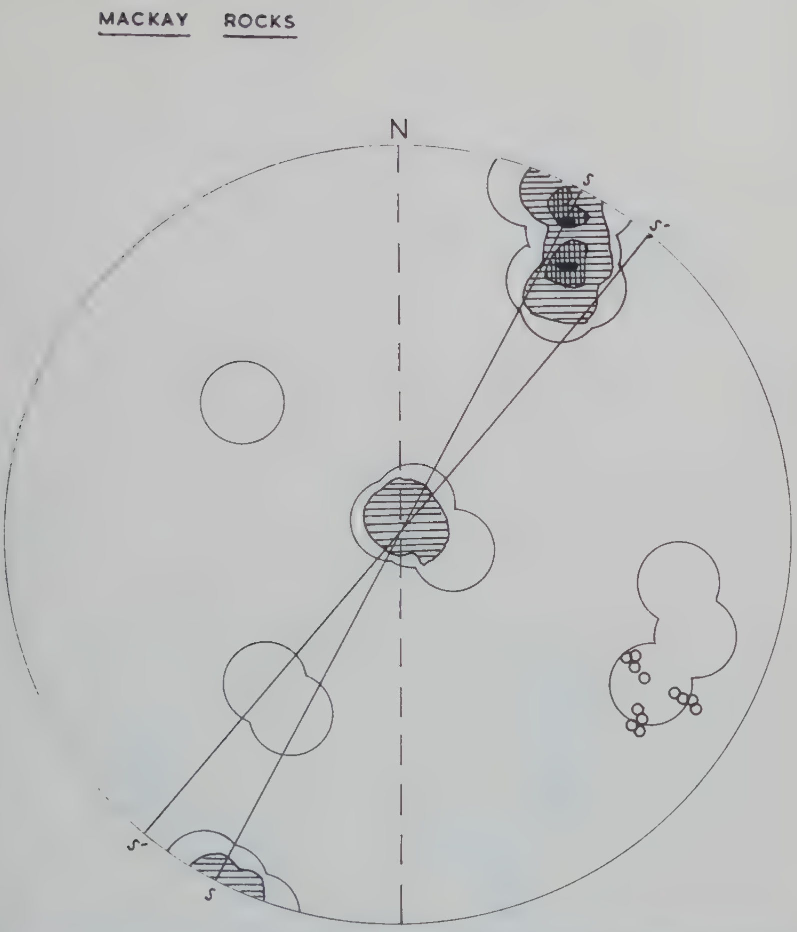

FIG. 4—Equal area projection of poles of 20 joints from Mackay Rocks contoured at 5%, 10%, and 20% per unit area, together with 12 poles of the bedding, S–S′ = range of strike of bedding.

Almer Ridge Joint Pattern

This is very similar to that of Mackay Rocks. Sixty-three major joints were recorded from Chl.2 and early biotite schists between Teichleman's Corner and the Almer Glacier and plotted on the equal area projection (fig. 5). The three minor 5% maxima of fig. 5 are from an area brecciated by faulting and are plainly discordant to the general structure.

Over the distance of three-quarters of a mile across the strike along Almer Ridge, the bedding strike varies by 33° about a mean of 025° dip 70°W and the main 10% concentration of joint poles lie in or near the zone

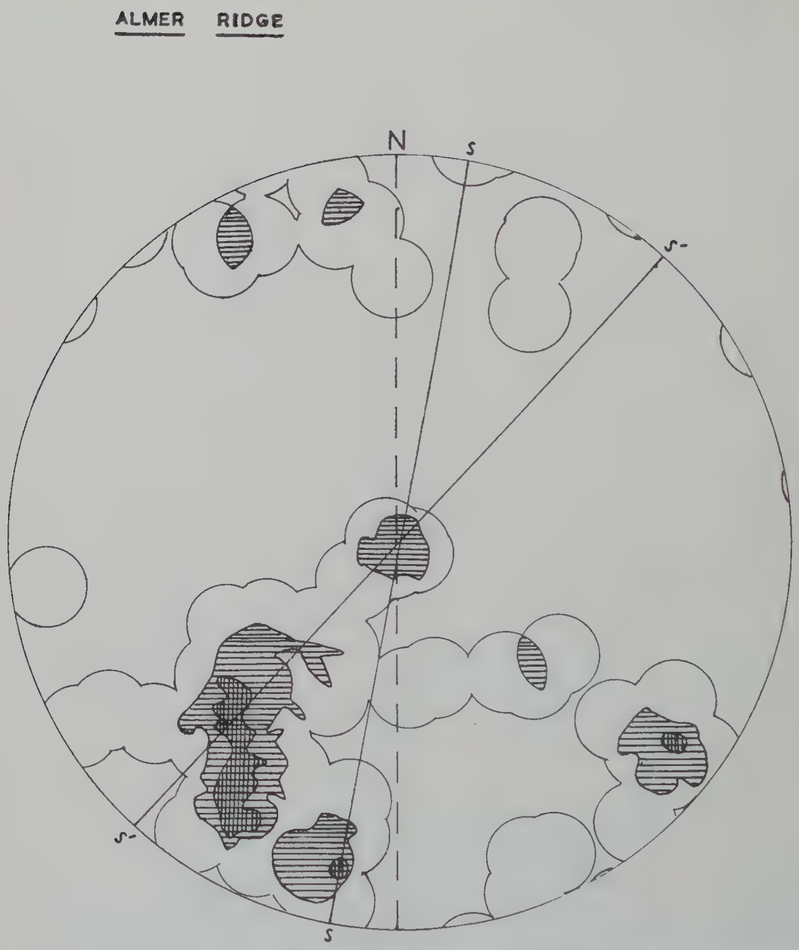

FIG. 5—Poles of 65 joints from Almer Ridge, contoured at 1·5%, 5%, and 10%, per unit area. S–S′ = range in strike of bedding

normal to the bedding. The axis normal to the joints has a plunge of about 35°S, parallel to the plunge of the major and minor folds a mile to the west, and so these are *a c* joints. The two other sub-maxima are bedding joints and sub-horizontal joints approximately perpendicular to the bedding but parallel to *b*.

Chancellor Ridge Joint Pattern

The projection of 54 joints and 17 bedding poles from the limbs of folds only, is shown in fig. 6. These are recorded from Chl. 2 and early biotite schists of Point A, Middle Deck and Chancellor Dome, Fox Glacier.

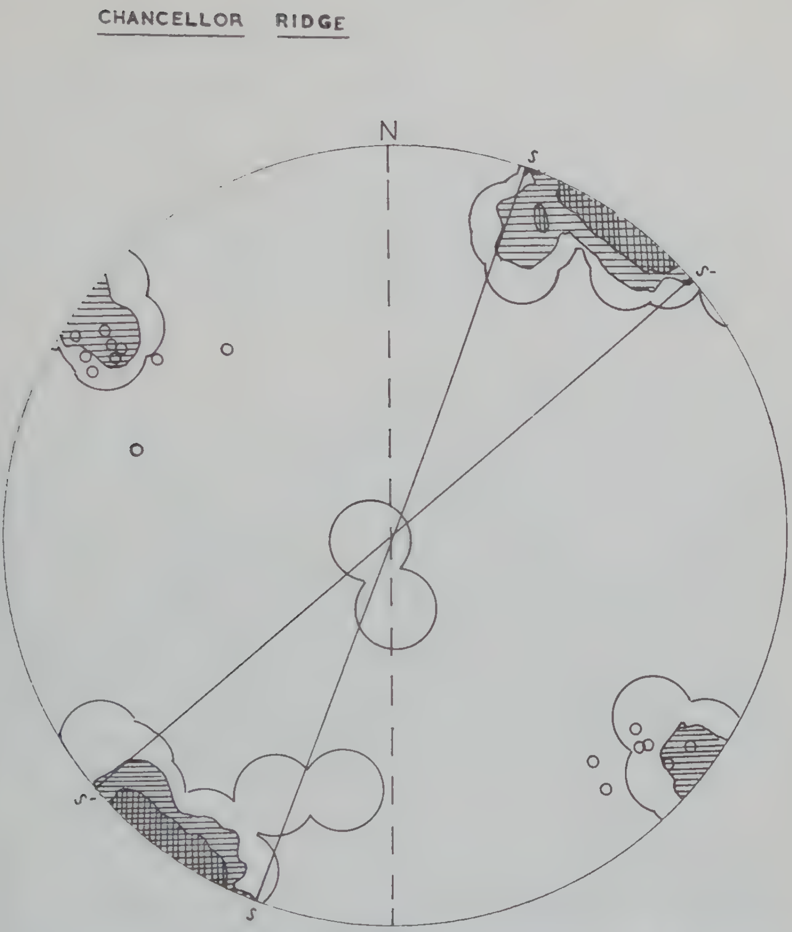

FIG. 6—Poles of 54 joints from Chancellor Ridge contoured at 2%, 8%, and 16%, together with 17 bedding poles.

The 16% concentration shows a joint strike of 126° with vertical dip. These are interpreted as being *a c* joints so that the *b* tectonic axis is horizontal, a point of considerable interest as the plunge of the lineation formed by the intersection of the axial plane and bedding cleavages is 32°S. The 10% concentration in part represents bedding joints and in part axial plane joints which are parallel to the axial-plane schistosity. The *a c* joints are less conspicuous in the higher grade schists to the west but joints in the axial planes of the folds become increasingly numerous and more closely spaced until in the proximity of the Alpine Fault the orientation of the bedding is completely masked by the schistosity and jointing.

QUARTZ VEINS

In the greywacke zone, anastomosing quartz veins are common in beds of indurated sandstone but without consistent orientation. In the Chl. 2 and higher grade schists there occur increasing numbers of quartz veins of three types.

1. Fine sheet-like veins parallel to the bedding.
2. Sheet-like quartz veins up to 5 in. thick, not parallel to the bedding.
3. Gash veins. These are lenses of quartz-feldspar, often as much as 12 in. thick but seldom more than 6 ft. long. Commonly the size is about 3 × 18 in. They are often curved in outline and terminate in several divaricating fingers.

Almer Ridge Quartz Vein Pattern

Fig. 7 shows the projection of the poles of 64 quartz veins from the Almer Ridge excluding those parallel to the bedding, together with 27 poles of the bedding. Fifteen of the quartz veins recorded are gash veins and show a fairly constant relationship to the bedding in that they strike roughly parallel to it but dip in the opposite sense, between 40° and 60°E. The poles of the sheet-like quartz veins show a single 20% concentration. They are of similar age to the gash veins and are interpreted as being the quartz filled joints of an earlier system. If the dominant concentration represents an *a c* joint pattern, then the fold axes during metamorphism were inclined at 35° to those of the present with a similar southward plunge of about 40°. No lineations or other evidence for a local change of fold axes were found, however. The equivalent schists of the Fox Glacier lie closer to a synclinal axis and the quartz veins are predominantly parallel to the axial plane of this and no check on the validity of the Almer Ridge pattern could be made.

FAULTS

The ice-scoured rocks of the glacial valleys plainly reveal fault contacts. Commonly a fine fault breccia, mylonite or soft gouge clay forms a band a few inches or feet in width, and a zone of intense brecciation of the hanging wall rock extends for a few feet or up to several hundred yards to one side. Faulting appears to be more intense in the western sector of the Alpine schists but with the exception of the two main glacial valleys the fault contacts of the western slopes of the Southern Alps are well concealed by forest and Pleistocene moraine.

The Alpine Fault

This major fault underthrusting the western escarpment of the Southern Alps was first recognised by Morgan (1908) in the neighbourhood of the Mikonui River in central Westland. Alpine schists were found to be thrust westward over Pleistocene moraine (p. 72) and Morgan considered the fault to be an underthrust of the type stated by Suess to be a common feature of fold mountains in general. Wellman and Willett (1942) traced the Alpine Fault southward as far as the coast of Fiordland and showed by the clock-

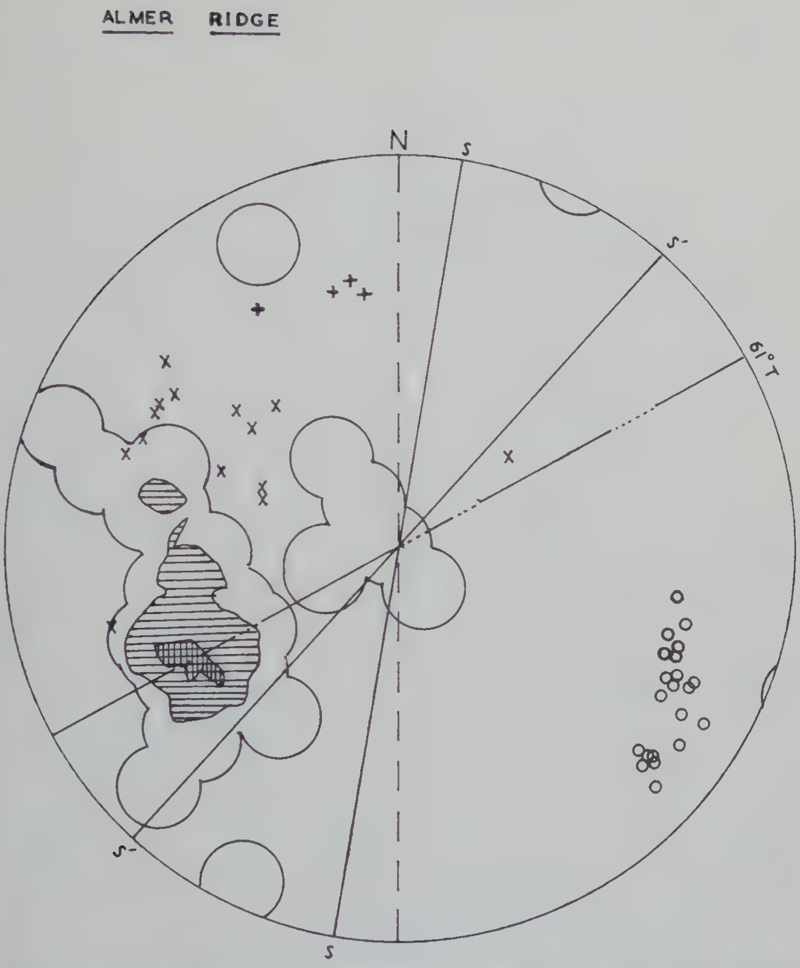

FIG. 7—Poles of 64 quartz veins contoured at 2.5%, 10%, and 20% per unit area. Highest concentration, 25%.

o = poles of bedding.
x = poles of quartz gash veins.
+ = poles of bedding disturbed by faulting.
S–S′ = range in strike of bedding.

wise displacement of river valleys that at least some of the displacement across the fault had been transcurrent. Wellman later suggested (1955, p. 38) that clockwise transcurrent displacement might have taken place over a distance of 300 miles, basing this hypothesis on the matching of pre-Cretaceous rocks on either side of the fault in western Otago and in Nelson. At the same time Wellman described an overthrust in the Alpine Fault at Paringa with a vertical displacement of not less than 1,500 ft. The observed dip of the fault was found to be 30° or less, as was the dip found by Bowen

(1954) for another outcrop of the overthrust in the Mt. Brown region south of the Taramakau River. To explain the apparent regularity of the strike of the fault and also its apparent transcurrent nature, Wellman (1955) suggested that the actual dip of the fault was nearly vertical and that the observable dip was merely the attitude of a slumped superficial gravity slide or nappe.

The Alpine fault outcrops at two localities in the Waiho-Weheka region. The outcrop in the Waikukupa valley has been previously noted (Wellman, 1953, p. 280) but not described, and the other outcrop at Stony Creek, 3 miles north of Waiho has not been previously reported. A third outcrop was seen in Gaunt Creek, a tributary of the Waitangi River, 9 miles to the north of Waiho, but this was not examined.

In both places visited, sheared and brecciated schist is thrust at a low angle westward over schist moraine belonging apparently to the last stage of the last glaciation.

At the Stony Creek outcrop the contact is subparallel to the schist lamination and dips 31°S, strike 090°. A crush zone extends 650 yd eastward and four more major faults render the rock unstable for another 400 yd beyond this. A total of 16 secondary faults from this zone were plotted stereographically and 9 of these fall in a zone about the most prominent schistosity which dips 40°SE. Some of these are splinter faults and some show downthrow to the west and are merely caused by slumping.

In the Waikukupa valley the alpine fault outcrops 270 yd above the bridge over the Hare Mare stream. Here the contact dips 27°E, strike 009° and has a clean-cut lower surface, the moraine boulders being bevelled and planed. About 30 ft of moraine is exposed surmounted by 12–15 in. of blue fault pug and an exposed 120 ft of brecciated schist set in a fault pug matrix. The crush zone is again about three-quarters of a mile wide and the secondary faults are roughly grouped about the schistosity at 040°/30°SE. On the south bank of the Waikukupa, breccia rises 600 ft above the river and extends at least 400 yd further west, indicating that the fault plane dips at a shallow angle. The moraine below the contact described above is compact and undisturbed and the dip of 27° is likely to be the true one and not merely due to slumping of a vertical contact.

The direction of nett displacement is not obvious but elevation of the Alpine schists due to the overthrusting cannot have been less than 600 ft since the last stage of the last glaciation.

Faults of the Franz Josef Valley

As shown in fig. 8 the faults of the Franz Josef form two distinct patterns of parallel faults: one east-west, and the other northwest-southeast. Some of the more important of these are described below.

Northwest-Southeast Faults

Park Rock Fault. This forms the southern side of five of the roches moutonnées lying in the Waiho river bed including Park Rock, Harpers Rock and Sentinel Rock and forms a prominent erosion nick up both valley walls. The strike is 140°, dip 65SW with the northwest side relatively upthrown.

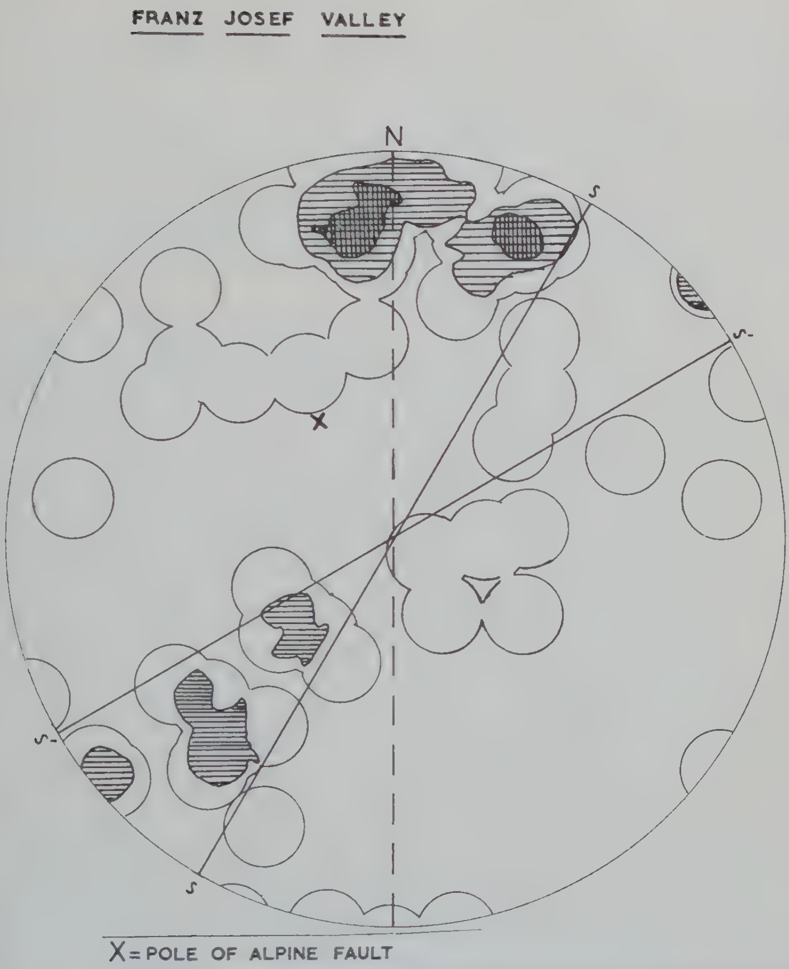

FIG. 8—Equal area projection of poles of 62 faults contoured at concentrations of 1·7%, 5%, and 10% per unit area
S–S′ = range in strike of bedding.

East-West Faults

These are most prominent along the lower Franz Josef Glacier where ice erosion has accentuated the fault scarps. The most important are:

Harpers Creek fault, strike 089°, dip 80°N. This fault is unusual in that both dip and brecciated zone are to the north of the fault plane. The fault line scarp rises almost 1,000 ft on the southern side and forms a prominent erosion nick known as Tower Saddle near Mt. Moltke. There is no obvious continuation of this fault on the eastern side of the glacier although extensive fault breccias are found at No Go Creek and south of the Almer Glacier.

Trident Creek Fault, strike 114°, dip 80°S. A fault line scarp rises 200 ft above the northern side of Trident Creek which cascades down the fault plane. On the east bank of the glacier the fault is multiple and in two cases slickensides show the northern wall to have been displaced relatively upwards and eastwards at an angle of 43°. This fault together with a similar fault lying 200 yd to the north, has elevated the bed of the Franz Josef Glacier and formed the lower icefall.

Rope Creek and Arch Creek flow along similar faults, both with a zone of breccia extending about 100 yd on the south side only. The slickensides of the Arch Creek Fault show the displacement of the north wall to be upwards and eastwards at an angle of 30°.

The Duck Creek Fault is also marked by an erosion scarp rising 200 ft on the northern side and four more east-west faults occurring within a quarter of a mile below Duck Creek are each marked by a small stream and a prominent re-entrant angle where the western upthrown side projects out into the river alluvium.

Summary of fault characteristics:

(a) The major east-west faults are spaced regularly at intervals of about 200 yd for a distance of at least six miles east of the Alpine Fault;
(b) Stress has not been relieved entirely by the formation of a few major faults and many minor faults showing displacements of a few inches each are found between them;
(c) The net slip direction of several of the east-west faults is clockwise along a line plunging 30°–40°W;
(d) With one exception it is the southern, hanging wall which has been brecciated;
(e) The NW–SE faults are parallel to the *a c* joint system.

Faults of the Fox Glacier Valley

Faulting has been rather more intense here than at Franz Josef, with the result that the valley sides are more unstable and have been reduced to flatter slopes and rockfalls and slips are more common. As a consequence the number of exposed fault contacts is less. The same two fault patterns found in the Franz Josef valley are found here, but the east-west faults are less prominent, probably because the valley lies in an approximate east-west direction.

The Cone Rock Fault

This has already been briefly described in an earlier publication (Lillie, Gunn and Robinson, 1957, p. 128). It may also be noted that slickensides on the face of Cone Rock show the relative displacement to be upward and north-westwards at an angle of 65° and that in Fork Creek three parallel fault surfaces show a combined displacement of about 100 ft. The fault is parallel to the *a c* joint system as is a similar fault, the north-eastern wall of which forms the opposite right bank of the lower Fox Glacier. The brecciated rock between has been selectively eroded to form the trough of the lower Fox Glacier. A similar series of faults along pre-existing *a c* joints line the walls of the Fox Glacier down the main icefall near Chancellor Hut. Zones of breccia connected with *a c* faults are found at Boyd Creek, White Creek, and Yellow Creek. Intersections of bedding faults and east-west faults form a zone of brecciation in Cone Rock itself and on the opposing side of the glacier, the western side having been displaced relatively upward. The central part of the Fox Glacier and the Victoria Glacier, both lie along a parallel series of small east-west faults seen both at the south end of Cone Rock and below the terminus of the Victoria Glacier.

Faults in the Alpine greywackes are common and are usually bedding faults overthrusting to the east. The displacement seldom appears to be more than a few feet.

The Tectonic Significance of the Fault Patterns

The faults striking northwest-southeast are, as stated previously, parallel to the *a c* joint system of the region. Shear stress later applied to the rock mass as a whole is likely to be relieved in part by slip along these pre-existing fractures. These faults have therefore no definite tectonic significance especially as the sense of displacement is not consistent. De Sitter (1956, pp. 160–172) points out that wrench faults normal to the fold axes are caused by difference in degree of folding.

Movement has taken place in the east-west faults at least as recently as the last movements of the Alpine fault, but they might well have been initiated during some stage of folding. Twenty-two major east-west faults were recorded between Cape Defiance and the upper swing bridge over the Waiho River. Cols in ridges, hot springs, and parallel streams might indicate more faults to the west.

The minor faults between the major faults have, in some cases, a total combined vertical displacement upwards of not less than 30 ft on the northern side. The amount of displacement of the major faults is unknown but to judge by the thickness of fault pug and breccia is likely to be appreciable. This means that a significant amount of elevation of the western boundary of the Alps has taken place by faulting alone, perhaps in the order of thousands of feet.

Cases in which diagonal faults intersect fold axes at angles of 30°–40° are not uncommon. Billings (1955, p. 213) cites the case of the Jura mountains where a single set of strike slip faults are developed at an angle of 30° to the normal of the fold axes. Here most of the east-west faults intersect the normal to the fold axes at an angle between 30° and 46° and may have a similar significance. It will be noted that the well-known Taramakau Fault is an east-west fault and also shows dextral displacement. The fold axes do not appear to have been markedly offset however and displacement along individual faults cannot have been more than a few hundred feet.

PETROFABRICS OF THE ALPINE SCHISTS

Preferred Orientation of Calcite

Lenses of marble up to 2 ft in thickness are found in the biotite schist at Rope Creek near the terminus of the Franz Josef Glacier and a specimen (No. 14834 of the Geology Dept. collection, University of Otago) was used for petrofabric analysis. The beds here dip 73° SE, strike 043° and are overturned, the axis of the Roberts Point Anticline is 300 yd farther eastward and dragfolds and lineations plunge 34° to 45° to the southwest.

In thin section the marble is seen to be composed of more than 70% calcite in grains which are more or less equidimensional in the plane of the bedding averaging 0·9 mm diameter, although they measure only 0·4 mm in the direction normal to the bedding. Rather more than half of the calcite grains show one or two sets of $(01\bar{1}2)$ twin lamellae, probably caused by minor post-metamorphic deformation. The remainder of the rock is made up of quartz, biotite, and muscovite, the micas all being oriented parallel to the bedding. The two equal area projections of the optic axes of two hundred grains each, both show a strong B-tectonite girdle pattern.

In fig. 9 constructed from a section cut parallel to the schistosity, the girdle axis plunges SW at 42°. Since this coincides with the regional plunge of fold axes at 35°–45°SW the girdle axis is the tectonic axis (*b*). As calcite optic

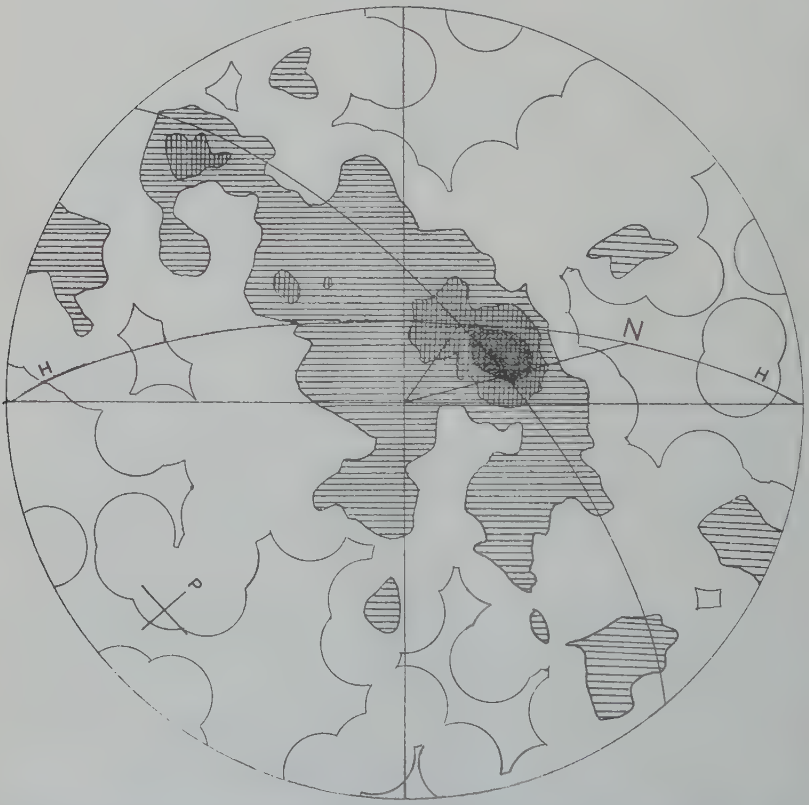

FIG. 9—Orientation diagram of 200 optic axes of calcite from Rope Creek marble, contoured at concentrations of 0·5%, 2%, 4%, and 6% per unit area, H–H = horizontal plane, N = true north. P = girdle axis.

axes tend to become oriented parallel to, or inclined at a high angle to the maximum principal stress (McIntyre and Turner, 1953, p. 227) the girdle would be compatible with rotation of the bedding during folding about the tectonic axis with a constant direction of compression. The subhorizontal maximum within the girdle is a common feature of such fabrics according to the same writers. It may indicate the main or last direction of compression.

A projection plotted from 200 optic axes of calcite in a section transverse to the bedding and in the vertical plane shows a similar strong girdle pattern and has a maximum at 325°T as in fig. 9.

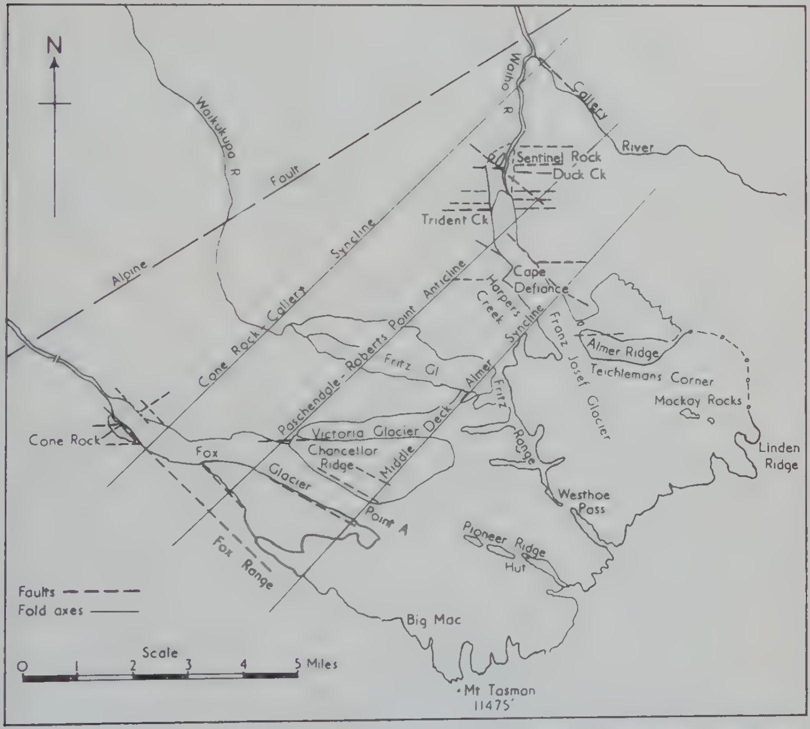

FIG. 10—Franz Josef - Fox Glacier region showing location of faults and fold axes.

Preferred Orientation of Biotite

A plot of 50 poles of cleavages of biotite from oligoclase-garnet-biotite schist from Park Rock shows a strong 8% concentration normal to the bedding and a weaker 4% concentration normal to the axial plane cleavage.

CONCLUSIONS

The rocks in the area under discussion have been subject to intense folding along northeast-southwest axes. During fold movements the maximum principal stress was directed at right angles to the fold axes. This resulted in:

(1) The formation of a system of joints parallel to the stress and normal to the tectonic axis (b);
(2) Differential slip between beds forming a bedding schistosity and strong dragfolding and crumpling;
(3) The formation of a vertical cleavage especially near the axes of folds this cleavage being normal to the principal stress;
(4) The initiation of a single set of shear joints or faults (the east-west faults) intersecting the normal to the fold axes at angles of about 40°;
(5) The development of a direction of preferred orientation of calcite grains in marble;
(6) The formation of a reverse fault underthrusting the north-western boundary of the folded schists (the Alpine Fault).

Renewed postmetamorphic underthrusting lead to the wedging-up of the western margin of the Alpine schists producing step-like displacements along pre-existing faults and joints.

The direction of the maximum principal stress during folding may be deduced according to the different lines of evidence mentioned above as follows:

Normal to fold axes: 306°–316°.
Principal joint strike: 300°–315°.
Preferred orientation of calcite: 325°.
Normal to Alpine fault: 328°.

ACKNOWLEDGMENTS

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THE GEOLOGY OF THE FLAT POINT AREA, EASTERN WAIRARAPA

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Summary

The stratigraphy and structure of the Flat Point area, eastern Wairarapa is described and illustrated by a map. Strata in upward succession are: Mokoiwian greywackes with spilites; Ngaterian to Piripauan graded-bedded sandstone, siltstone and conglomerate; Haumurian and Teurian brown siltstone with glauconitic sandstone; Lower Dannevirke limestone and greensand; Upper Dannevirke and Bortonian bentonitic siltstone and greensand; Pareora and Southland alternating sandstone and siltstone.

Dolerite intrusions that cut Ngaterian and Raukumara strata have alkaline affinities and are co-magmatic with teschenites described to the north.

The strata are strongly folded along north-north east fold axes, and are disrupted by parallel faults.

Introduction

The Flat Point area (Fig. 1) is about 25 miles south-east of Carterton, Wairarapa, and occupies parts of sheets N 162, N 163, N 166 (N.S.M.S. 1). The area is broken and hilly, but cleared of forest. Access is fairly good, a metal road from Masterton reaches Te Wharau, where it branches to the Kaiwhata River and Flat Point.

Approximately two months were spent in the field, the larger streams and the coast being traversed. The whole area is covered by air photographs and these were of considerable help in mapping, but do not show any structural detail. Most of the rocks slump badly and none is well exposed. Graded bedding was used for determining the order of superposition of strata, which is not always evident from dips alone.

Previous Geological Work

W. A. McKay (1899) divided the sedimentary rocks from the Kaiwhata River to Glenburn into four main divisions, which now have little significance. Sollas and McKay (1906) described the ophitic hypersthene dyke rocks collected by McKay from near Flat Point. King (1930) described marine and river terraces. No detailed geological study or map had been made previously.

Stratigraphy

Stratigraphic subdivision is based on fossils – mainly *Inoceramus* and Foraminifera – and is essentially expressed in terms of existing Cretaceous (Wellman, 1959) and Tertiary (Finlay and Marwick, 1947) stages.

The stages and formations recognised in the area are set out in fig. 2.

N.Z. J. Geol. Geophys. 3 : 309–320

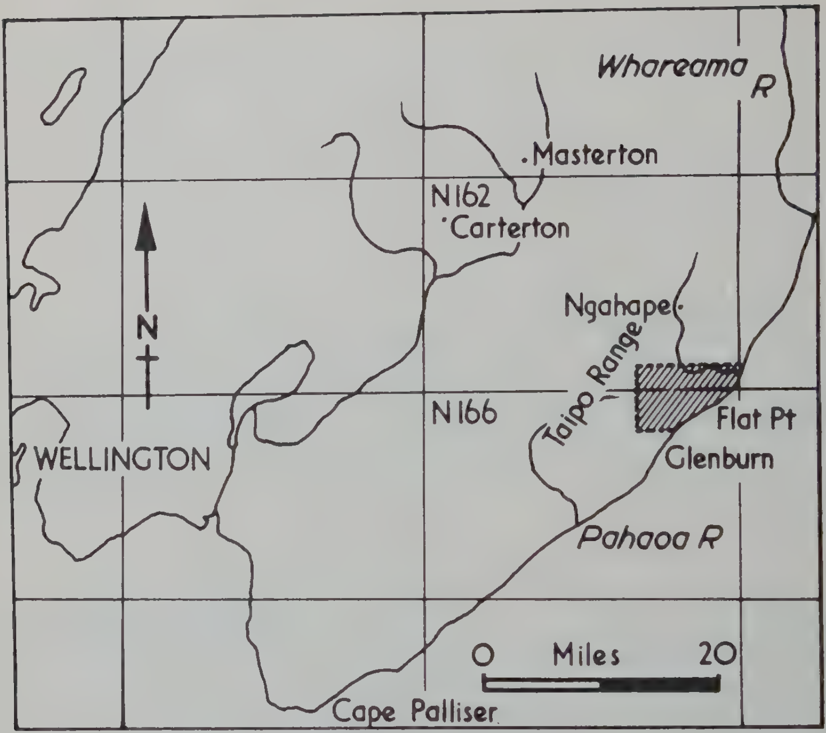

FIG. 1—Locality map of the Flat Point Area.

Taitai Series

Mokoiwian Stage

Mokoiwian strata are confined to the western part of the area and cover about 2 square miles in the upper valley of the Pahaoa River. They consist of argillite, greywacke, and thin conglomerate with interbedded spilitic lavas. The rocks are intensely sheared and poorly exposed. *Inoceramus warakius* is relatively abundant within calcareous concretions in the argillite.

The best section in an unnamed northwest-flowing tributary of the Pahaoa River (N162/315317)* comprises a westerly younging, overturned sequence of sheared, blue-grey and green-grey argillites with calcareous concretions and interbedded, dark grey, graded-bedded, greywacke overlain near the Pahaoa River by unfossiliferous dark grey, graded-bedded sandstone and thinly bedded, carbonaceous siltstone. Calcareous concretions contain abundant large fragments and small complete shells of *Inoceramus warakius* (N162/782)†.

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*Grid reference. †N.Z. fossil number, see Appendix.

Boulders of conglomerate are present 20 chains south-west of Anderson Trig Station. In Anderson Stream a 50 ft wide, 60° westerly dipping, volcanic sequence is faulted against the Whangai Shale. It is composed of dark-green and red spilites with amygdules filled with calcite and interbedded with jaspers and red argillites. Similar spilites are poorly exposed in a stream south of Anderson Trig Station.

The following composite section is exposed in tributaries of the Pahaoa River.

Mokoiwi Section, Anderson Trig Station

	Ft
Graded-bedded sandstone and thin carbonaceous siltstone	+ 400
Crushed dark siltstone with interbedded, graded-bedded sandstone, calcareous concretions with *Inoceramus warakius*	600
Crushed dark siltstone with interbedded sandstone	*c.* 800
Graded-bedded sandstone and siltstone with minor thin conglomerate beds	*c.* 200
Crushed dark siltstone with interbedded sandstone	*c.* 1,000
Red and green spilitic lavas, jaspers, and red argillites	50

The Mokoiwian on its western side is unconformably overlain by Lillburnian strata; to the east it is in fault contact with the Whangai Shale.

Similar strata with rare *Inoceramus* fragments (probably *I. warakius*) are exposed in a small faulted inlier 1 mile south-east of Anderson Trig Station. Spilites and jaspers show at fault contacts to the north and east. Near the junction of Flat Point and Craigielea roads a small inlier of grey sandstone, fine conglomerate, and dark-grey argillite with *Inoceramus* fragments is probably also Mokoiwian in age.

Similar rocks occur in other parts of the Wairarapa, one of the best sections being near the mouth of the Pahaoa River, where dark crushed mudstone containing poorly preserved *Inoceramus* (probably *I. warakius*) is overlain by massive sandstone (Waterhouse *in* Wellman 1959). Thick greywacke sandstone and associated conglomerates with igneous pebbles, in the Taipo Range (McBeath 1950) resemble the Taitai Sandstone of the Raukumara Peninsula (Wellman 1959) and may overlie the Mokoiwian of the valley of the upper Pahaoa River.

Clarence Series

Ngaterian Stage

The rocks of the Ngaterian Stage consist of more than 2,000 ft of graded-bedded greywacke sandstone and argillite, with minor conglomerate bands and concretionary layers. No strata intermediate in age between the Mokoiwian and Ngaterian are known, but the base of the Ngaterian is unfossiliferous and may be older than mapped.

Two belts of Ngaterian strata have been mapped. The western belt is cut off by Fault IV (Fig. 3) in the Arawhata Valley and extends north for 4 miles to the Kaiwhata River. It is about half a mile wide and the beds dip west from 60° to 120°. It is conformably overlain by Arowhanan sediments to the west and is in fault contact with Haumurian to Bortonian strata along Fault IV to the east. The eastern belt extends from the Huatokitoki Stream to the Kaiwhata River. It is variable in width owing to faulting, and the beds dip east from 60° to 110°. To the west it is in contact with Haumurian to Bortonian sediments along Fault V; to the east it is conformably overlain by Arowhanan strata near Unu Unu Stream, but is faulted against Raukumara strata at its northern and southern extremities. Dolerite sills intrude the southern part of both belts.

In both belts the lower 1,500 ft of strata are unfossiliferous, the upper 500 ft are fossiliferous. The unfossiliferous part consists of thick graded-bedded greywacke sandstone alternating with indurated dark argillite and may be pre-Ngaterian. A few conglomerate bands, up to 6 ft thick, with rounded pebbles of greywacke, argillite and concretionary fragments, occur at the base of several of the greywacke beds. The fossiliferous part is composed of indurated, thinly bedded, alternating dark-grey fine sandstone and blue-grey argillite with calcareous concretions. Fragments of *I. fyfei,* and I. *hakarius* occur in the argillite and in the concretions. Layers of *Inoceramus* fragments form limestone bands up to 2 in. thick at a few places.

Raukumara Series

Three belts of Raukumara beds have been mapped. The western belt extends from Shephard Trig Station to Kaiwhata River road. It is about half a mile wide and the beds dip west at 50° to 60°. It is conformably overlain by Mata sediments on the west and in fault contact with Haumurian to Dannevirke sediments to the east. Dolerite intrusions are present in its most southern part. The southern end of the middle belt is cut off by Fault IV in the middle Arawhata Valley, and the belt extends north for 4 miles to the Kaiwhata Valley. It is about three quarters of a mile wide and the beds generally dip west from 60° to 70°. It is overlain by Piripauan sediments to the west and overlies the Ngaterian strata. Several dolerite sills, up to 20 ft thick, are present in its southern part.

The eastern belt, also cut off by a fault at its southern end, extends north for about 4 miles to be cut off by Fault V. It dips east generally from 40° to 70°, and is about three quarters of a mile wide at its widest point. The beds conformably underlie Piripauan sediments to the east and conformably overlie Ngaterian sediments to the west.

Inoceramus is fairly abundant, the four key species, *I. rangitira, I. bicorrugatus, I. opeteus,* and *I. nukeus* having been recognised. They occur in their established stratigraphic order (Wellman 1959), with the possible exception of *I. bicorrugatus* which was found only in boulders. Except in the western belt *Inoceramus* is sufficiently abundant to define the position of the three Raukumara stages fairly accurately. The distribution of the species is given in Appendix A; those found in place are shown on the map by the initial letter of their specific names. The Arowhanan Stage is defined by *I. rangitira,* the Teratan Stage by *I. opeteus* and *I. nukeus.*

The Raukumara Series is about 2,500 ft thick, the Arowhanan and Mangaotanean being each about 500 ft thick, and the Teratan about 1,500 ft thick. The beds, uniform throughout, consist of graded-bedded, laminated and cross-laminated sandstone and siltstone with a few mud-flake and pebble conglomerates. The colour of the sediments changes from dark grey in the lower part to light grey and green-grey in the upper part. The pebble con-

Stratigraphic Sequence of the Flat Point Area, Eastern Wairarapa

New Zealand Series	Stages	Map Symbol	Lithology	Thickness ft
	Recent	F	Blown sands, alluvium, etc.	
Hawera			Marine terrace veneer . . . Plio-Pleistocene Unconformity	4
Southland	Lillburnian	Sl	Alternating sandstone and siltstone . basal conglomerate. Unconformity at base . .	1,100
Southland	Clifdenian Altonian	Sa–c	Graded-bedded sandstone and siltstone, basal conglomerate . . .	2,200
Pareora	Waitakian? Hutchinsonian Otaian	P	Graded-bedded sandstone and siltstone, glauconitic, unconformity at base? .	700
Arnold Dannevirke	Bortonian to Heretaungan	Dh–Ab	Bentonitic siltstone, graded-bedded glauconitic sandstone Huatokitoki Bentonite . . .	c.300
Dannevirke	Mangaorapan Waipawan	Dw–m	Graded-bedded greensand, limestone Kaiwhata Limestone . .	c.300
Mata	Teurian Haumurian	Mh–T	Brown siltstone, glauconitic sandstone Whangai Shale . . .	c.1,500
Mata	Piripauan	Mp	Graded-bedded, quartzose sandstone, siltstone, and conglomerate, glauconitic	400
Raukumara	Teratan	Rt	Graded-bedded sandstone, siltstone and conglomerate, carbonaceous . .	1,500
Raukumara	Mangaorapan	Rm	As above	500
Raukumara	Arowhanan	Ra	Graded-bedded sandstone, siltstone, and thin conglomerate . . .	500
Clarence	Ngaterian	Cn	As above . . . . Lower Clarence not exposed in area?	2,000
Taitai			Korangan-Taitai Sandstone exposed west of the area	
	Mokoiwian	Um	Crushed siltstone with graded-bedded sandstone, spilites at base . .	c.3,000

Fig. 2—Map legend and stratigraphic sequence for Fig. 3.

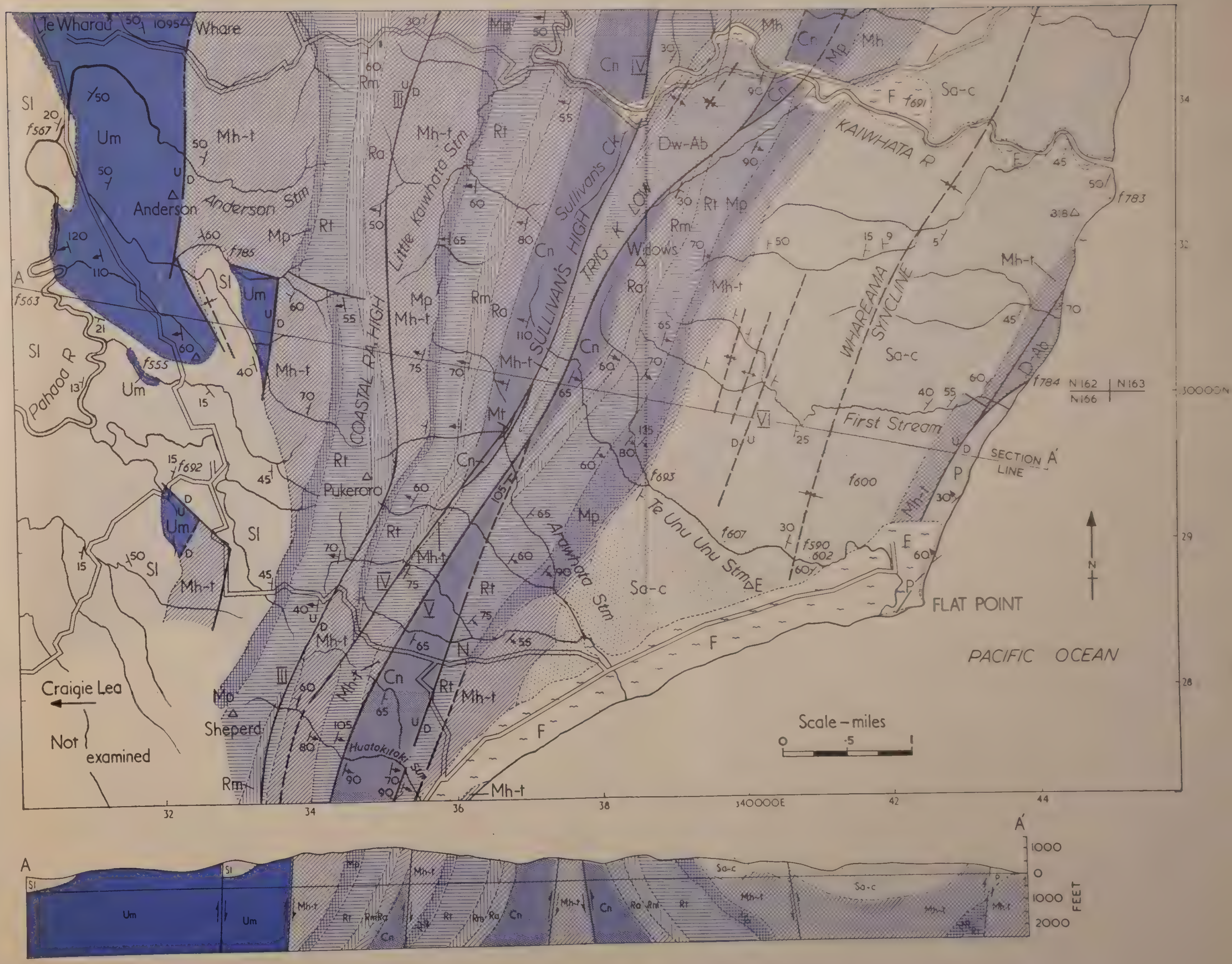

Fig. 3—Geological map of the Flat Point Area, Eastern Wairarapa

glomerates are up to 9 ft thick and are composed of rounded pebbles of greywacke, argillite, quartz, and derived concretions. None of the pebbles is particularly distinctive and they could have been derived from any of the older greywacke and argillite groups of New Zealand.

Mata Series

The three Raukumara belts already described are conformably overlain by Mata beds. The western belt is about a mile wide. It is faulted against Mokoiwi beds to the west and is largely covered by unconformably overlying Lillburnian beds to the south. The middle belt is faulted against Raukumara beds along Fault III. The eastern belt is unconformably overlain by Altonian beds to the east. The beds dip steeply to the west in the two western belts and to the east in the eastern belt. Within the three belts the upper part of the Mata Series is not exposed; it crops out along the Trig K Low where it conformably underlies the Dannevirke Series, but it cannot be clearly related to the lower and middle beds exposed in the three belts, and the full thickness of the Mata Series is uncertain. Upper Mata beds also occur near the coast between Flat Point and the Kaiwhata River, but they are poorly exposed and their thickness and stratigraphic relations remain uncertain.

The most complete section through the lower and middle part of the Mata Series is exposed in the northern tributary of the Arawhata River, 10 ch north of, and parallel to, the line of cross section. It lies near the centre of the middle belt and dips uniformly west at 70° to 75°.

Mata Series, Northern tributary of Arawhata River

	Ft
Brown micaceous siltstone with laminae and thin lenses of greensand	*c.* 600
Graded-bedded glauconitic sandstone and siltstone with a 3 ft contorted band in middle	50
Dark-grey slightly graded siliceous siltstone with large spherical concretions	250
Dark-grey graded-bedded fine sandstone and siltstone contorted in places, with thin greensand dykes	70
Yellow-grey graded-bedded sandstone and brown siltstone, carbonaceous	120
Yellow-grey graded-bedded sandstone and thin grey siltstone, some glauconitic and thin conglomerates	60
Grey and green-grey graded-bedded sandstone and siltstone	150

The section consists essentially of an upper part 900 ft thick, composed mainly of brown siltstone, and a lower part, 400 ft thick, of graded-bedded sandstone and siltstone. Fragments of the key Piripauan fossil *Inoceramus pacificus* were found in similar graded-bedded strata at Little Kaiwhata Stream. The overlying siliceous and micaceous siltstone contains scattered pyrites and produces a yellow efflorescence on protected surfaces. It is thus similar to the widespread Whangai Shale of Haumurian to Teurian age (Wellman 1955), with which it is correlated. Several microfaunal samples were collected in order to define the Piripauan-Haumurian and Haumurian-Teurian contacts but all proved to be barren of Foraminifera. The Piripauan-Haumurian contact is considered to lie at the top of the graded-bedded sandstone, about 400 ft above the base of the Piripauan; the Haumurian and Teurian stages are grouped together on the map. The only other fossils seen are flattened worm tubes and fragments of thin shells in the upper part of the series at the coast near Glenburn and western flank of the Coastal Range (N162, 343313, and 327322).

Dannevirke and Arnold Series

Wanstead Group

Because of their poor exposures and disrupted nature the Waipawan to Bortonian strata were mapped together as the Wanstead Group. The name Wanstead Formation was introduced by Quennel and Brown (1936) for basal Tertiary beds in the Dannevirke Subdivision. It includes several distinct lithologies and is more appropriately termed the Wanstead Group (Waterhouse and Bradley 1957).

The group can be subdivided into the: Huatokitoki Bentonite of Mangaorapan to Bortonian age and the Kaiwhata Limestone of Waipawan to Mangaorapan age.

Waipawan to Mangaorapan, Kaiwhata Limestone

The Kaiwhata Limestone, about 300 ft thick, is exposed in the infaulted Trig K Low, extending from the Kaiwhata River to the Huatokitoki Stream. It is folded, faulted, and severely crushed. The following section exposed in two tributaries of the Kaiwhata River north of K4 Trig Station (N162, 321347) conformably overlies the Whangai Shale and is faulted against Ngaterian strata.

Kaiwhata Limestone, tributaries Kaiwhata River

	Ft
Greensand and glauconitic sandstone	+ 75
Massive, grey-white flinty limestone	20
Alternating grey marl and flinty limestone with interbedded greensand	17
Alternating grey marl and flinty limestone with minor greensand	*c.* 30
Sequence uncertain	30
Greensand	20
Grey-white, flinty limestone	10
Quartzose sandstone, greensand, and siltstone	100

Most of the glaucontitic sandstones are graded-bedded and have flute casts and worm tracks as sole markings. The limestone thickens to the south and is 50 ft thick in the Huatokitoki Stream. Some of the limestone beds contain elongated mud flakes, contorted laminated siltstone layers and thin greensand bands.

Several microfaunal samples were collected, but all proved to be barren. A sample collected by Laing, probably from the top of the formation, is Waipawan to Mangaorapan (N 166/608). Its relationship to the Whangai Shale and the Huatokitoki Bentonite suggests a Waipawan and Mangaorapan age for the Kaiwhata Limestone.

The Mungaroa, Manurewa (Waterhouse and Bradley 1957), and the Kaiwhata Limestone probably represent the northward extension of the lower part of the Amuri Limestone of Marlborough.

Mangaorapan to Bortonian, Huatokitoki Bentonite

At most places in the infaulted Trig K Low the Kaiwhata Limestone is overlain by about 300 ft of green, grey, and reddish bentonitic siltstones and interbedded glauconitic sandstones. Similar slumped and poorly exposed deposits occur at the coast between the Kaiwhata River and First Creek. One- to 10-inch bands of reddish, green and grey bentonitic siltstones are interbedded with glauconitic sandstones up to 10 inches thick. The bentonitic siltstone shows characteristic "coral" weathering described by Macpherson (1952, p. 278). White plates of calcite, weathered from the bentonite, litter outcrops and slips. The glauconitic sandstone is graded-bedded and has flute casts and worm tracks as sole markings.

A microfaunal sample from coastal beds is Bortonian (N162/784); other samples proved to be barren. Similar strata near the mouth of the Pahaoa River (N166/656) contain a Mangaorapan to Heretaungan fauna (Rishworth 1953). The Huatokitoki Bentonite is thus probably Mangaorapan to Bortonian in age.

Pareora Series

Approximately 700 ft of alternating sandstone and siltstone of probable Pareora age are clearly exposed on the coastal platform 10 to 40 chains north of Flat Point. On its western side they are faulted against the Whangai Shale and the Huatokitoki Bentonite.

Pareora Series north of Flat Point

	Ft
Grey, graded-bedded sandstone and siltstone with interbedded massive mudstones up to 30 ft thick	350
Grey, graded-bedded sandstone and siltstone with thin greensand dykes near base, N166/707, 708	220
Massive light green mudstone	10
Alternating, grey sandstone and siltstone, glauconitic in places	+ 100

The sandstone is laminated, cross-laminated, and convolute-laminated, graded bedding is general, and mud flakes and grit occur at the base of some beds. A variable amount of glauconite is present in all sandstones. Interstratal folds, contortions, and thin greensand dykes occur in the basal 200 ft.

Microfaunal samples 200 ft (N166/707) and 300 ft (N166/708) above the base are younger than Duntroonian and older than Altonian and are probably Pareora Series in age (Mr P. Vella, pers. comm.).

The relation between these deposits and the Huatokitoki Bentonite and Altonian strata is obscure.

Southland Series

Altonian and Clifdenian

Grey, graded-bedded sandstone and siltstone fill a 2-mile wide, north-northeast-trending syncline that extends from the mouth of the Arawhata River into the Whareama district (Pick 1955). The beds are about 2,200 ft thick and rest unconformably on older rocks.

The lower 200 ft is composed of conglomerate, grit, sandstone, and siltstone. Conglomerate beds, prominent on the western flank of the Whareama Syncline, contain pebbles of greensand, sandstone, limestone, concretionary fragments, greywacke, mudstone and igneous rocks that were largely derived from Upper Cretaceous and Lower Tertiary strata. On the coastal platform south of the Kaiwhata River mouth, the basal zone contains thick contorted siltstones with inclusions of brown siltstone, greensand, and conglomerate.

The middle 300 ft, best exposed in sea cliffs south of the Kaiwhata River, is composed of layers of graded-bedded sandstone and siltstone up to 3 ft thick, with laminated, cross-laminated and convolute-laminated structures, outlined by carbonaceous material.

The top consists of graded-bedded sandstone and siltstone, from 1 to 12 inches thick. Carbonaceous material is fairly common. Fine shell fragments occur at the base of several sandstones.

Microfaunas from the basal beds (N162/783, N166/693) have few species, but are not older than Altonian. Higher samples are Altonian (N166/599,/600,/0607) and Clifdenian (N162/691).

The Lower Southland strata of the Flat Point area are similar to those of the Whareama district that were considered by Pick (1955) to have been transported by turbidity currents, and except for the basal conglomerate and grits, appear to have been deposited in the same way.

Lillburnian

Lillburnian beds cover much of the western part of the area. They unconformably overlie Upper and Lower Cretaceous sediments and are conformably overlain by Waiauan beds to the west. The total thickness is 1,100 ft (McBeath 1950), but only the lower and middle parts are exposed in the area described here.

The basal 200 ft consists of massive conglomerate, sandstone, and siltstone. The conglomerate contains rounded to sub-rounded pebbles of Whangai Shale, greensand, limestone, greywacke, and argillite derived from Upper and Lower Cretaceous strata. Worn molluscan shells are abundant, the following species predominating:

Dosinia sp.
Zeacolpus sp.
Struthiolaria callosa
Polinices huttoni
Baryspira sp.
Dentalium solidum

To the south-west the basal beds are massive yellow-brown sandstone with beds and lenses of coquina limestone.

The middle part is composed of alternating sandstones and siltstones, with fine conglomerates and shell grits at the base of most sandstones. The grits at the base of the sandstone bands are absent in the highest horizons mapped, and siltstone predominates.

Microfaunal samples (N162/785, N166/692) from the basal beds are Lillburnian to Waiauan. Macrofaunas collected by McBeath (1950) (N162/549–55,/561–64,/567) are Lillburnian.

Lillburnian strata do not rest on Lower Southland strata in this area, as they are reported to do by Pick (1955) in the Whareama district. Altonian sediments are recorded south of Ngahape, and near Te Wharau, and were apparently eroded off the western part of the area prior to the deposition of the Lillburnian.

Hawera Series

Extensive flat remnants of a marine terrace (King 1930) flank the coast from Glenburn to the Whareama area. They are about 500 ft above sea level in the southern part and gradually decrease in height to the north. The terrace extends inland for 2½ miles, and is covered by a few feet of poorly consolidated sandstone and fine conglomerate.

The height and degree of dissection of the terrace suggest a Hawera age.

Holocene

Post-Hawera deposits consist of uplifted beach deposits, blown sands, river terrace deposits, and alluvium.

Raised beach deposits, up to 12 ft thick occur along the coast from Flat Point to south of the Kaiwhata River. They rest on a wave-cut platform, 2 to 3 ft above present-day high-water level and consist of well bedded sandstone, siltstone, and fine conglomerate. Molluscan shells belonging to living species occur at the base of these deposits.

Blown sands cover a coastal strip up to half a mile wide from Flat Point to Glenburn. At its inner edge at Unu Unu Stream Mollusca similar to those on the coast are embedded in the sand. Alluvial fans deposited by coastal streams cover large areas.

River terraces were not studied in detail. Low-level terraces, at a height of 6 and 8 ft, are present in all coastal streams. Higher terraces, both aggradational and degradational, are present in all major streams.

Igneous Rocks

Upper Cretaceous dolerite intrusives

Basic sills from a few inches to 20 ft thick occur at several horizons within Ngaterian and Raukumara strata. A few well exposed contacts show both underlying and overlying sediments to have been thermally metamorphosed for up to 10 inches. No tuffs, pillow lavas, or other definite volcanics were seen.

Petrography: Macroscopically the rocks range from fairly coarse types, grey-green in colour, with abundant crystals of augite, feldspar, and iron ore; to fine-grained varieties, green-black in colour. They are holocrystalline with occasional sub-ophitic texture and are considerably altered. The main mineral constituents are basic andesine and augite; accompanied by pleochroic biotite and hornblende, partly altered to green chloritic; minerals; magnetite and ilmenite with leucoxene; apatite. A varying amount of calcite, and indeterminable carbonates, analcite, and alkali feldspar, probably orthoclase, are present in some of the sills. Quartz has been recorded in one specimen. The mineral composition and texture indicate that the rocks are altered dolerites with alkaline affinities.

The dyke rocks from near Flat Point described by Sollas and McKay (1906, pp. 153–4) as "ophitic hypersthene andesite", are identical to the dolerites described above (see also Hutton 1943, p. 367). They are similar to, and probably the same age as, the teschenites described by Brown (1943) and Hutton (1943) from the Brocken Range - Ngahape area, 5 miles to the north. Hutton (1943) considered a variolitic pillow basalt near Ngahape, interbedded in the Whangai Shale (Wellman 1956) to be co-magmatic with the teschenites. Haumurian to Teurian is thus the most probable age for the igneous intrusion.

Mokoiwian Spilites

These rocks are very fine grained, grey-green to red in colour, and contain amygdules of calcite. In some places they are intensely sheared the small albite-oligoclase crystals being oriented so as to give the rocks a variolitic appearance. Augite present in noteworthy amount is fragmental and occasionally bent. Accessory minerals include iron ores and chlorite.

STRUCTURE

The strata are strongly folded and faulted, and crop out as steeply dipping north-northeast-striking belts. Major folds in Cretaceous strata are obscure, probably because of faulting-out of one of the limbs. Most major faults are based on stratigraphy, but where exposed, are represented by crush zones. The main north-northeast faults, marked on the map with roman numerals, are described from west to east. Faults I and II, between the Whangai Shale on the east and Taitai strata on the west, have throws of at least 5,000 ft. Fault III is based solely on stratigraphy, the inferred throw being at least 2,000 ft. Fault IV has wide crush zones extending from the Kaiwhata River to the Arawhata River, but dies out further south. A 30 ft wide faulted outlier of Whangai Shale, in Clarentian strata, is exposed in the Arawhata River, a few chains west of the main fault. At its northern end the western side is upthrown at least 2,500 ft, but the throw decreases further south. Fault V has wide crush zones, mainly in bentonitic siltstones. In contrast to the previously described faults the eastern side is upthrown, the throw being at least 3,000 ft. Fault VI is mapped to explain basal Altonian conglomerate in First Creek; the northern and southern extensions of the fault are hypothetical. Fault VII separates westward-dipping Whangai Shale from westward-dipping Bortonian and Pareora strata. The western side of the fault is upthrown at least 1,300 ft. The faults are shown as steeply dipping reverse faults on the cross section, but there is no definite evidence for their nature or attitude. The numerous northwest-, west-, and southwest-striking faults, most of which are not mapped, have little structural importance. No evidence for recent faulting was seen.

Slumping and Clastic Dykes

Intricately contorted zones with numerous inclusions of greensand and sandstone were seen in all thick sections of the Whangai Shale, and are particularly abundant in the eastern part of the district. Greensand dykes are usually restricted to the less disturbed Whangai Shale, but at Glenburn they cut across contorted beds and are younger than the contortions. Thin greensand dykes and contorted inclusions of siltstone were observed in the Kaiwhata Limestone at Kaiwhata River. Waterhouse and Bradley (1957) have described slumping in beds of this age at White Rock, 40 miles to the south. Pareora strata, exposed along the coast north of Flat Point, have small interstratal folds resembling convolute lamination, and a few thin greensand dykes. Small folds in Altonian strata, similar to slump folds described by Pick (1955) from the Whareama area are present in alternating strata south of the mouth of the Kaiwhata River. They are directly overlain by contorted siltstones, with inclusions of sandstone, greensand, conglomerate and brown siltstone.

ACKNOWLEDGMENTS

The work was carried out in 1958 at the Victoria University of Wellington, in partial fulfilment of the requirements for a Master of Science degree in Geology. The author wishes to thank Mr J. Bradley and especially Dr H. W. Wellman, of Victoria University, for their criticism of the manuscript and for help in the field, and Mr W. R. Lauder, of Victoria University, for assistance in petrology. Mr G. H. Scott, N.Z. Geological Survey, and Mr P. Vella, Victoria University of Wellington, kindly identified microfaunal samples. Grateful acknowledgments are made to residents of the district for their hospitality and assistance in transport.

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APPENDIX

Inoceramus Fossil Localities

Fossil Sheet No.	Locality	Species	Remarks
N162/782	315317	*I. warakius*	Calc. concretions
*N162/791	315330	*I. warakius*	In situ
*N162/	324310	*I. warakius*	In situ
*N162/	333316	*I. warakius ?*	Fragments in slip
N162/787	390326	*I. fyfei*	In stream boulders
N162/789	382308	*I. fyfei*	In situ
N162/793	382336	*I. fyfei*	In stream boulders
*N162/800	372314	*I. fyfei*	In stream boulders
N166/694	366295	*I. fyfei I. hakarius*	In situ
N166/695	353252	*I. fyfei*	In situ
*N166/696	369288	*I. fyfei I. hakarius*	In stream boulders
N166/697	355268	*I. fyfei*	In situ
N166/698	359270	*I. fyfei ?*	In situ
N162/790	382307	*I. rangitira ?*	In situ
N162/797	370327	*I. rangitira ?*	In situ
N162/798	353328	*I. rangitira*	In situ
*N162/799	370317	*I. rangitira*	In stream boulders
*N162/801	365304	*I. rangitira*	In stream boulders
*N166/704	365295	*I. rangitira*	In situ
*N162/794	367330	*I. bicorrugatus*	In stream boulders
*N162/801	365304	*I. bicorrugatus*	In stream boulders
N166/702	370282	*I. bicorrugatus*	In stream boulders
*N166/703	331245	*I. bicorrugatus*	In stream boulders
N162/788	388307	*I. opeteus*	In situ
N162/795	368342	*I. opeteus ?*	In stream boulders
*N166/700	368285	*I. opeteus*	In situ
N166/706	362293	*I. opeteus*	In situ
N166/691	386302	*I. nukeus*	In situ
N166/699	373275	*I. nukeus*	In situ
*N166/701	362267	*I. nukeus*	In situ
N162/786	399333	*I. nukeus*	In situ
N162/792	373353	*I. nukeus*	In situ
N162/796	370344	*I. nukeus*	In situ
*N162/	359322	*I. pacificus*	In situ
*N162/	361293	*I. pacificus*	In stream boulders

*Not collected.

NOTE ON TERTIARY ROCKS IN THE UPPER GREY VALLEY, WESTLAND (S45)

By B. D. WEBBY, New Zealand Geological Survey, Department of Scientific and Industrial Research, Greymouth

(Received for publication, 5 October 1959)

Summary

Three isolated remnants of Tertiary strata crop out in the Upper Grey Valley. In Clarke River and at Staircase Creek there is Landon limestone and sandy limestone; between Clarke River and Allan Water an infaulted sequence includes, in upward succession, coal measures, Whaingaroan calcareous siltstone and mudstone, and unconformably overlying breccias and grits, and Waitotaran-Nukumaruan Coal measures.

INTRODUCTION

The rocks of the Upper Grey Valley (fig. 1) consist of: (1) pre-Mesozoic basement of intensely fractured greywacke and argillite, gneiss, foliated gneissic granite, and porphyritic granite; (2) Tertiary strata including coal measures, limestone, greensand, siltstone, mudstone, breccias, and grits; (3) late Pleistocene glacial and fluvioglacial deposits. Three isolated remnants of Tertiary rocks are preserved, along north-north-east trending faults which cut the basement rocks. One is on the north-east side of the Clarke River one mile upstream from its mouth, a second at the mouth of Staircase Creek, and the third on the banks of the Grey River between Clarke River and Allan Water. The late Pleistocene deposits include the products of two valley glaciations. Ice advanced down the Grey Valley as far as the mouth of the Clarke River during the first valley glaciation, leaving remnants of brown, slightly weathered moraine in Frank Creek, and on the track between Clarke River and Frank Creek. Beyond the ice limit, outwash gravels were built up to about 250 ft above the present valley floor. In the second valley glaciation ice advanced to Gentle Annie Gorge, within 4 miles of Staircase Creek, and outwash gravels were built up to about 150 ft above the present valley floor beyond the ice limit.

The Upper Grey Valley area is situated between the Grey - Inangahua Depression to the west, and the mountainous, granite-gneiss complex of Mt. Elliot and Victoria Range to the east. The present eastern margin of the gently dipping upper Tertiary sediments of the Grey - Inangahua Depression is one mile west of Allan Water (Dibble and Suggate, 1956, fig. 2).

In the Reefton district, to the north of the Upper Grey Valley, extensive areas of Arnold and Landon strata are exposed, mainly in the vicinity of Garvey Creek, Capleston, and Merrijigs (Suggate, 1957). Also, in the Alexander River, an infaulted strip of Tertiary rocks occurs (Gage, 1948, p. 10).

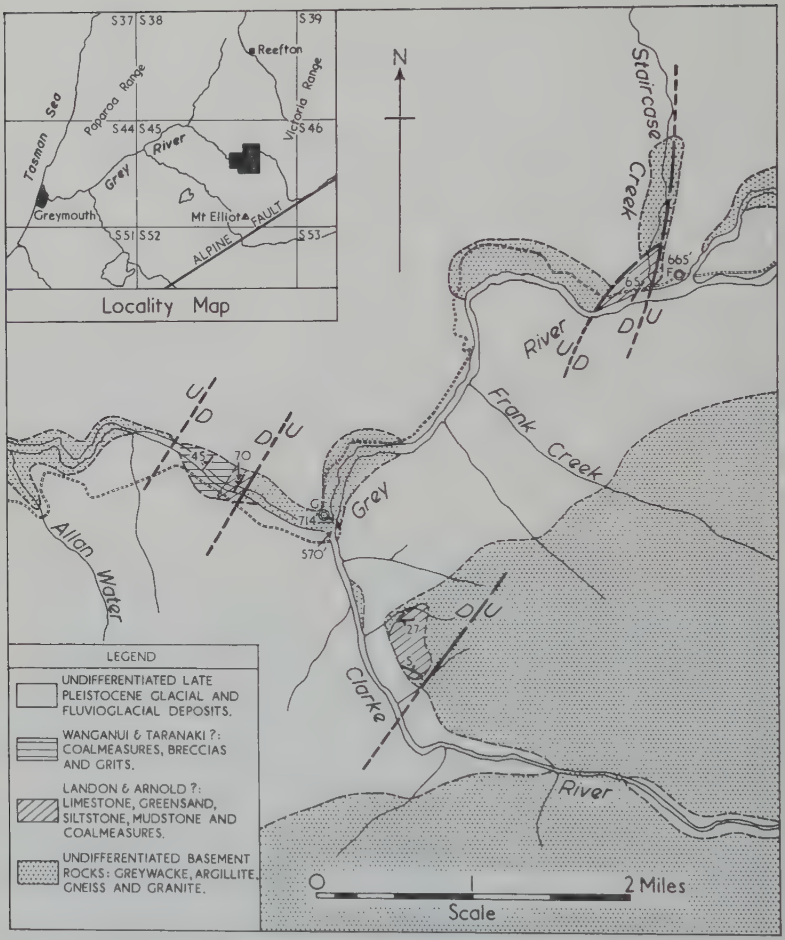
S37
S38
S39
Reefton
Paparoa Range
Victoria Range
Tasman Sea
S44
S45
S46
River
Grey
Greymouth
Mt Elliot
FAULT
ALPINE
S51
S52
S53
Locality Map
N
Staircase Creek
665'
F
65
River
U
D
D
U
Frank Creek
U
D
D
U
45
70
G
714
570'
Grey
Allan Water
D
U
27
5
Clarke
River
LEGEND
UNDIFFERENTIATED LATE PLEISTOCENE GLACIAL AND FLUVIOGLACIAL DEPOSITS.
WANGANUI & TARANAKI ?: COALMEASURES, BRECCIAS AND GRITS.
LANDON & ARNOLD ?: LIMESTONE, GREENSAND, SILTSTONE, MUDSTONE AND COALMEASURES.
UNDIFFERENTIATED BASEMENT ROCKS: GREYWACKE, ARGILLITE, GNEISS AND GRANITE.
0
1
2 Miles
Scale

FIG. 1—Geological and locality maps of the Upper Grey Valley.

STRATIGRAPHY AND STRUCTURE

The isolated remnant of Tertiary rocks on the north-east side of the Clarke River consists of 150 ft of bedded, slightly glauconitic, buff-coloured limestone and sandy limestone. It is gently dipping, and lies unconformably on basement greywacke and gneiss. To the east the limestone is bounded by a fault, which has a north-north-east trend, and a crush zone $1\frac{1}{2}$ chains wide, exposed in the underlying, intensely fractured greywacke. Calcareous algae, possibly *Lithothamnion*, are seen in the limestone, but are not abundant. A sample (S45/522, F13716) from near the base of the limestone yielded a Whaingaroan microfauna (Mr G. H. Scott, *pers. comm.*).

Greenish yellow arenaceous sandstone dipping 45° to the west, and containing quartz veins and indeterminate casts of small shells, was observed at the mouth of Staircase Creek by Haast (1861, p. 101). Morgan (1921, p. 22) considered these to be quartzites of the Reefton Series. However, Gage (1948, p. 10) in a more detailed account, described the rocks as, "limestone, arenaceous limestone containing irregular patches and lenses of arkosic fragments, and a true arkose with calcareous cement, passing upwards into moderately soft sandstone. . . . The limestone contains the calcareous alga *Lithothamnion*, a typical Tertiary fossil, and a large foraminifera identified by Dr Finlay as *Amphistegina lessonii*, an inflated shallow-water form that characterised the Miocene (Awamoan-Hutchinsonian)". A sample (S45/523, F13884) from towards the base of the limestone contained very poorly preserved foraminifera, which indicated a Kaiatan to lower Southland age (Mr G. H. Scott, *pers. comm.*). This determination, and the lithological similarities with the Whaingaroan limestone in Clarke River, suggest a Landon not a Miocene age (Gage, *loc. cit.*), for the Staircase Creek Limestone.

On the banks of the Grey River, downstream from the mouth of the Clarke River, an infaulted strip of Tertiary rocks shows the following section:

Series and Stages	(Fault)	Thickness (ft)
Wanganui (Waitotaran-Nukumaruan)	Grits and carbonaceous siltstones with thin coal seams (coal measures)	100
	(Not seen)	600
? Taranaki	Compacted, poorly interbedded breccias and grits composed of granite, and less abundant greywacke fragments. Granite boulders up to 2 ft long (Unconformity)	1,000
	Weathered laminated greensand	3
Landon (Whaingaroan)	Grey calcareous siltstone and mudstone with rare lamellibranchs (Sample S45/519, F13713, Whaingaroan)	200
	Greensand and glauconitic calcareous siltstone with calcareous algae, possibly Lithothamnion	25
Arnold (? Runangan)	Grits and carbonaceous siltstones with thin coal seams (coal measures) (Sedimentary contact) Leached greywacke basement	70

The vertical, faulted, sedimentary contact at the base of the Tertiary sequence strikes north-north-east. Downstream (west) from the contact the beds dip west-north-west between 70° and vertical below the intra-Tertiary unconformity. Immediately above this unconformity the beds dip west-north-west between 35° and 45°. The coal measures at the top of the sequence also dip steeply, and are almost certainly faulted against basement gneisses, although the contact cannot be seen. A sample (S45/519, F13713), from the grey calcareous siltstone and mudstone 100 ft below the weathered laminated greensand, contained a good Whaingaroan microfauna (Mr G. H. Scott, *pers. comm.*). It is assumed that the Landon limestone at this locality was eroded during the time represented by the intra-Tertiary unconformity. Also, a sample (S45/517, L1595) of carbonaceous siltstone from the coal measures at the top of the sequence was examined by Dr W. F. Harris. He reported that the flora has a Quaternary aspect, and is judged to be of an age near the Waitotaran - Nukumaruan boundary. It is not a cold flora suggesting the climates of the lower Nukumaruan glaciations and may therefore represent a mild interstadial or the cooling which led up to the first glaciation and brought the Pliocene (Waitotaran) to a close. It seems reasonable to regard these vertically dipping uppermost Waitotaran coal measures as equivalent to the gently dipping upper Pliocene coal measures known to rest directly on basement rocks and underlie "Old Man Gravels" in the Grey Valley, west of Allan Water. The difference of dip of the upper Pliocene coal measures gives some indication of the intense differential earth movements in the upper Grey Valley during mid-Pleistocene time. Judged by their degree of induration, the breccias and grits, which underlie the upper Pliocene coal measures, are possibly Taranaki in age.

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AGE RELATIONS OF THE DUNEDIN VOLCANIC COMPLEX AND SOME PALEOGEOGRAPHIC IMPLICATIONS—PART I

By D. S. Coombs, A. J. R. White, and D. Hamilton, Geology Department, University of Otago

(With an Appendix by Anne Scott, New Zealand Geological Survey, Department of Scientific and Industrial Research, Lower Hutt.)

(Received for publication, 2 November 1959)

Abstract

Marine inliers at Dowling Bay – Waipuna Bay and Blanket Bay on the northern shores of Otago Harbour are of Waiauan, or at youngest, early Tongaporutuan age. They contain mixed basaltic and trachytic tuffs of the Initial Eruptive Phase at a number of horizons among their upper members, thus fixing the date of initiation of volcanic activity. The beds pass conformably upwards into basaltic tuffs followed by flow-rocks of the First Major Eruptive Phase. The sediments comprise massive and laminated micaceous sandstones, calcareous sandstones and fetid limestones (Dowling Bay Limestone) as well as the interbedded water-lain tuffs. All these, up to the top of the highest sandy layer at Waipuna Bay, are grouped as the Waipuna Bay Formation. They are derived mostly from the Otago schists, but contain minor microcline of granitic or high-grade metamorphic parentage. Outward dips of the sediments and overlying volcanics are ascribed to doming by magmatic intrusions.

Introduction

In the course of a paleomagnetic study of Cenozoic volcanic rocks in New Zealand (Coombs and Hatherton, 1959), it became desirable to determine the age relations of the Dunedin Volcanic Complex more precisely than had hitherto been attempted. A detailed study of the Dowling Bay – Waipuna Bay and Blanket Bay sedimentary inliers (locality map, fig. 1) has provided new information on the initiation of volcanism and is here reported. Microfloral, paleomagnetic and geomorphic evidence, together with paleogeographic implications, will be discussed in a separate part of this paper.

The late Professor W. N. Benson (1940, fig. 1; 1942, p. 90; and MS) considered the Dowling Bay Limestone of Waiauan age (Upper Middle Miocene) to be the youngest Tertiary marine sediment in the Dunedin district and to be overlain unconformably by the volcanic sequence. Park (1904) also stressed the unconformable relationship. On the west side of Dunedin, e.g., St. Clair cliffs, the surface of contact is clearly an erosion surface. Benson (1935; 1942, p. 85) invoked a pre-volcanic Late Tertiary peneplanation and mainly for physiographic reasons he considered the volcanic activity to have occurred in Pliocene times, thus allowing time in the Upper Miocene and possibly earlier Pliocene for the development of the peneplain. Oliver (1936, p. 286) favoured an age "at least as old as Lower Pliocene" for the flora of the Kaikorai Leaf Beds which were considered by Benson to be intermediate in age between his First and Second Major Eruptive Phases. Later, in view of work by Couper (1953 A, B) on microfloras Benson (1959, MS.) accepted a probable Upper Miocene age for the com-

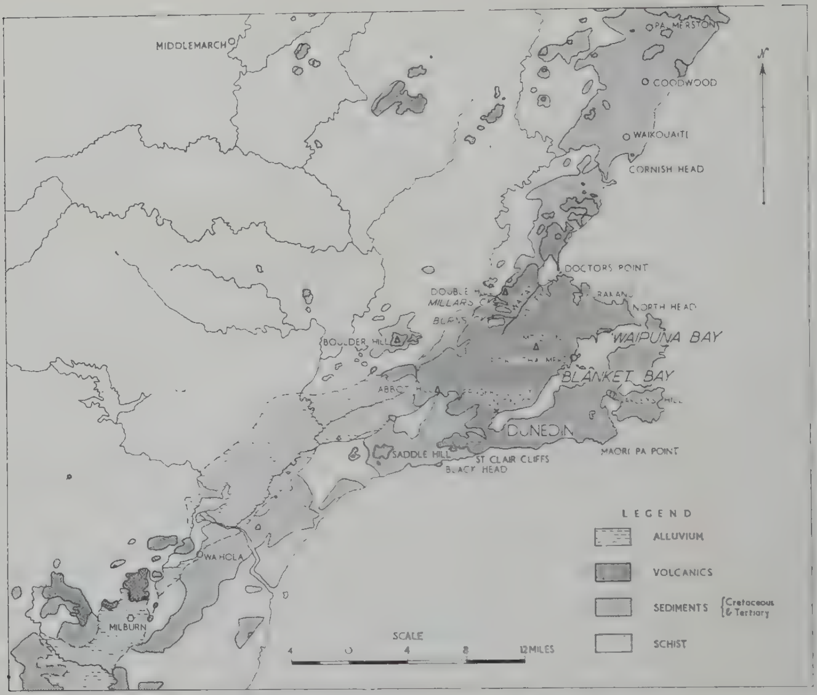

FIG. 1—Locality map of Dunedin district showing distribution of volcanic and sedimentary rocks (after Benson 1946).

mencement of volcanicity but realised that the short time interval for the supposed peneplanation became a serious anomaly.

The present work results in the demonstration of a Waiauan or possibly early Tongaporutuan age (late Middle or early Upper Miocene) for the initiation of the volcanic activity. A conformable relationship is established between the youngest marine sediments and early volcanic products in the Dowling Bay and Blanket Bay areas of Otago Harbour where tuffs are interbedded with fossiliferous shallow-water sediments and are followed by lavas. These facts require modification of the Late Tertiary peneplain concept.

The writers are grateful to the Director and several officers of the New Zealand Geological Survey for paleontological age determinations: Mrs Anne Scott (mollusca), Mr N. de B. Hornibrook, and Mr G. H. Scott (foraminifera), and Dr C. A. Fleming for helpful comments. In addition our debt to the late Professor Benson and his writings cannot be overstressed.

THE DOWLING BAY - WAIPUNA BAY INLIER

Previous Work

Blair (1879) reported the occurrence of limestone at Dowling Bay. The molluscan fauna was described by Finlay and McDowall (1923) who published a map of the area prepared by Benson. Subsequently, Finlay and Marwick (1948) and Benson (MS) gave the age as Waiauan on foraminiferal evidence.* Finlay and McDowall show sandstone overlying limestone, and state (p. 106) that another sandstone, correlated with the Caversham Sandstone, underlies the limestone. At Waipuna Bay, the supposed Caversham Sandstone is shown as being directly overlain by a volcanic sequence which overlaps westwards on to the upper sandstone.

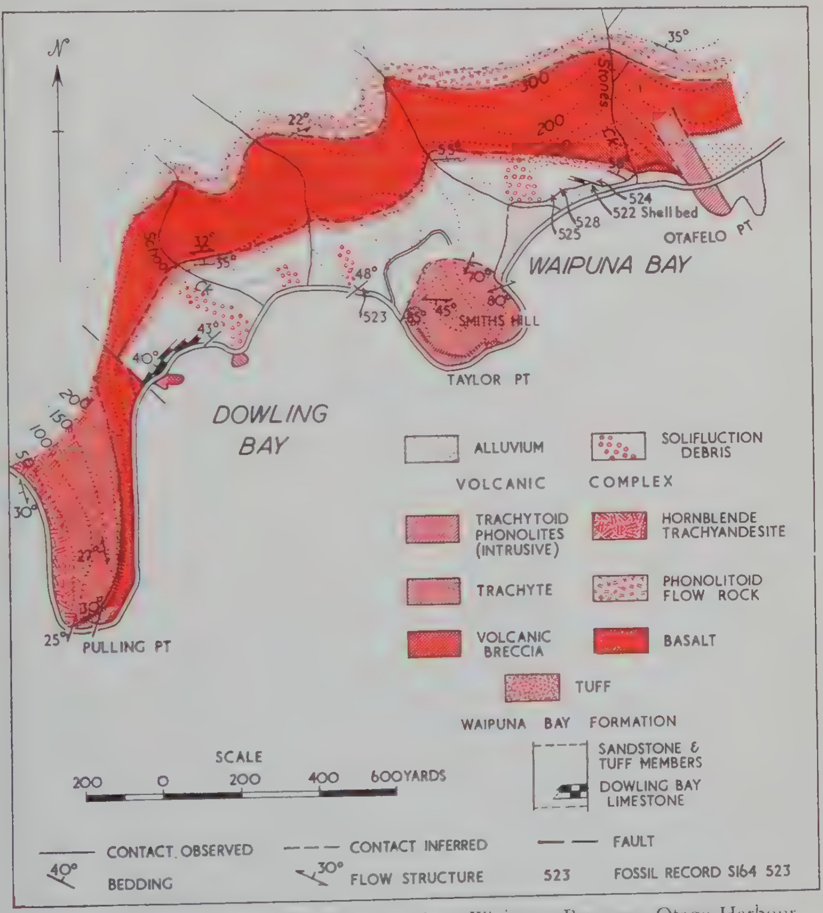

FIG. 2—Geological map of the Dowling Bay - Waipuna Bay area, Otago Harbour.

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*Mr N. de B. Hornibrook (pers. comm.) has re-examined a number of collections (S164/485–490) said to come from Dowling Bay Limestone and interbedded mudstones at Dowling Bay and Waipuna Bay and originally sent by Benson to Finlay. Mr Hornibrook reports that some of the faunules are datable only approximately but a few contain the pelagic *Orbulina suturalis* (Clifdenian upwards) and two have *Rectobolivina maoria* (top Altonian upwards). Several contain specimens referable to *Loxostomum truncatum*, index for Waiauan and Lower Tongaporutuan. There is a possibility of confusing this species with a long-ranging undescribed species of *Bolivina*, but specimens from S164/488 are so thickened and truncated that Mr Hornibrook considers that they can only be identified as *L. truncatum*.

Disregarding one faunule, which is suspected to come from a misplaced sample, the microfaunas thus indicate an Upper Southland Series age as being the most likely.

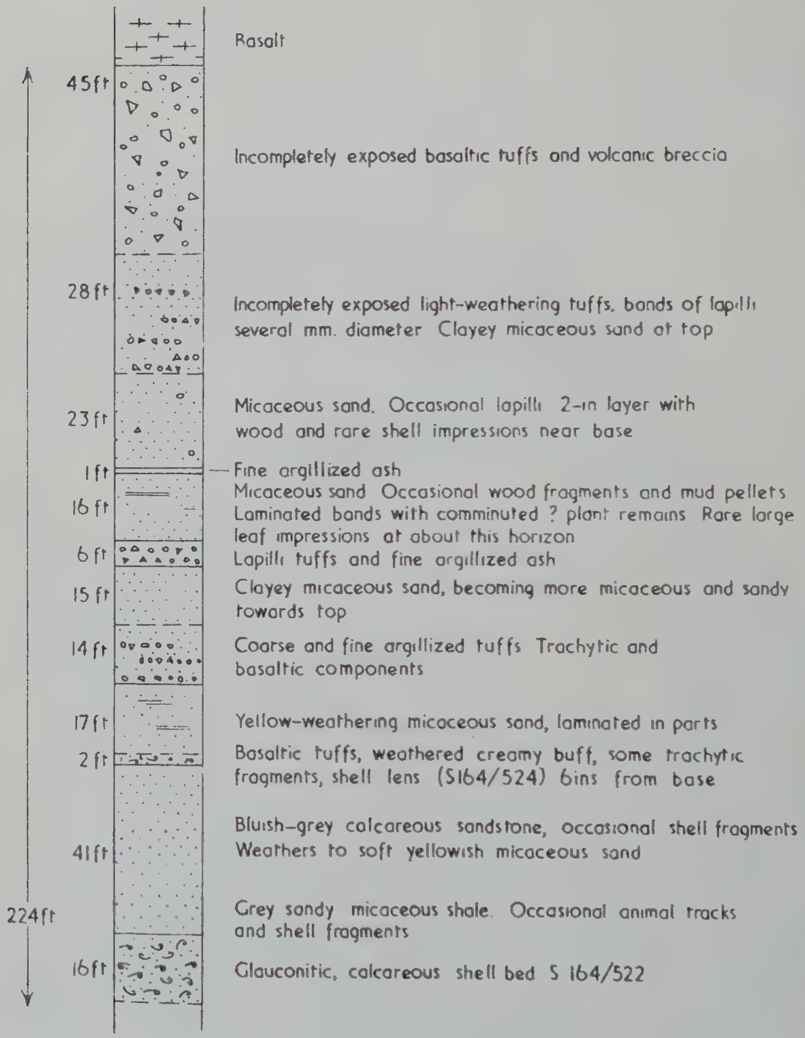
Basalt
45ft
Incompletely exposed basaltic tuffs and volcanic breccia
28ft
Incompletely exposed light-weathering tuffs, bands of lapilli several mm. diameter Clayey micaceous sand at top
23ft
Micaceous sand. Occasional lapilli 2-in layer with wood and rare shell impressions near base
1ft
Fine argillized ash
16ft
Micaceous sand Occasional wood fragments and mud pellets Laminated bands with comminuted ? plant remains Rare large leaf impressions at about this horizon
6ft
Lapilli tuffs and fine argillized ash
15ft
Clayey micaceous sand, becoming more micaceous and sandy towards top
14ft
Coarse and fine argillized tuffs Trachytic and basaltic components
17ft
Yellow-weathering micaceous sand, laminated in parts
2ft
Basaltic tuffs, weathered creamy buff, some trachytic fragments, shell lens (S164/524) 6ins from base
41ft
Bluish-grey calcareous sandstone, occasional shell fragments Weathers to soft yellowish micaceous sand
224ft
Grey sandy micaceous shale. Occasional animal tracks and shell fragments
16ft
Glauconitic, calcareous shell bed S 164/522

FIG. 3—Section on east side, Waipuna Bay.

New Observations

The sandstones at Waipuna Bay are well exposed in an abandoned sea cliff a chain or so north of the road where they strike north of west and dip 50°–58°N. A revised geological map is shown in fig. 2. Measured sections are shown in figs. 3 and 4. The upper 224 ft (fig. 3) down to and including a 16-ft sandy glauconitic shell bed, are exposed almost continuously on the west side of a tree-lined indentation on the cliff about 100 yd west of Stones Creek. The sandstone is locally calcareous, but it is deeply leached and it is

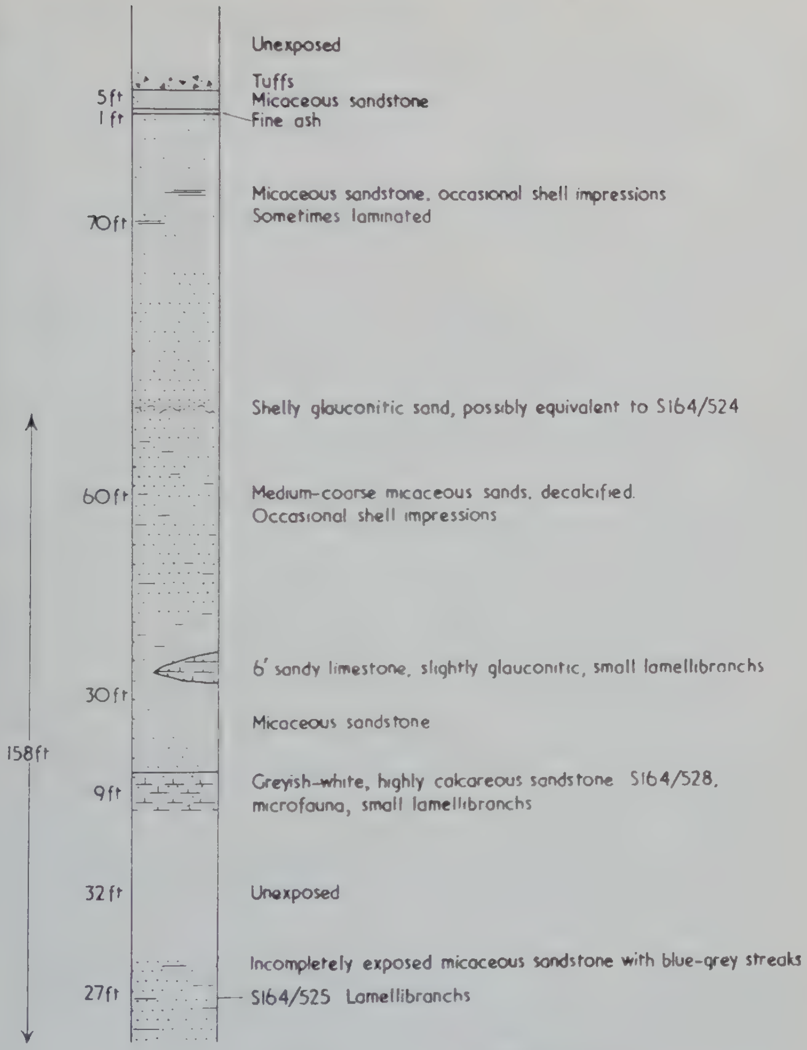
Unexposed
Tuffs
5 ft
Micaceous sandstone
1 ft
Fine ash
Micaceous sandstone, occasional shell impressions
Sometimes laminated
70 ft
Shelly glauconitic sand, possibly equivalent to S164/524
60 ft
Medium-coarse micaceous sands, decalcified.
Occasional shell impressions
6′ sandy limestone, slightly glauconitic, small lamellibranchs
30 ft
Micaceous sandstone
158 ft
9 ft
Greyish-white, highly calcareous sandstone S164/528,
microfauna, small lamellibranchs
32 ft
Unexposed
Incompletely exposed micaceous sandstone with blue-grey streaks
27 ft
S164/525 Lamellibranchs

FIG. 4—Section on west side, Waipuna Bay.

difficult to judge how calcareous most of it was originally. The section consists essentially of interbedded micaceous sandstones and tuffs. The tuffs become more abundant upwards, and except perhaps for the upper members appear to be water-sorted.

The shell bed (S164/522) yielded a fauna of Waiauan or early Tongaporutuan age (Appendix A). More mollusca were collected from a 2-ft ash bed (S164/524) 41 ft higher and others were noted as rare impressions at other horizons. Occasional large leaf impressions were found towards the top of the section.

A longer but poorly exposed section was measured 100 yd to the west (fig. 4). The main shell bed becomes increasingly difficult to trace westwards from the first section, but its probable position is suggested by a poorly exposed fossiliferous, glauconitic sand. About 90 ft beneath this is a rather hard, highly calcareous sandstone (S164/528) yielding a good Waiauan microfauna very similar to Dowling Bay collections (Mr G. H. Scott, pers. comm.), as well as some small indeterminate lamellibranchs. Forty-five feet below S164/528, a fine micaceous sandstone (S164/525) yielded casts of *Neilo sublaevis* Marwick, indicating a Southland Series, post-Awamoan age (Appendix A).

Outcrops are lacking on the slopes immediately north-west of Smiths Hill but a small thickness of calcareous sandstones and blue-grey micaceous mudstones is exposed under bouldery soliflual deposits in a road-side bank at the extreme east of Dowling Bay. Our mapping suggests these are the lowest beds exposed in the inlier. We consider that the lithology is very different from that of the Caversham Sandstone which Finlay and McDowall considered (1923, pp. 106, 112) to be exposed here. One specimen of *Chlamys* was collected and a microfauna (S164/523B) proved to be in most respects comparable with, though poorer than, previous collections from the Dowling Bay Limestone. It contains the distinctive Waiauan to lower Tongaporutuan form, *Loxostomum truncatum* (Mr G. H. Scott, pers. comm.).

The faunal evidence thus confirms that the sandstones at both Dowling Bay and Waipuna Bay are distinct from the Caversham Sandstone, the youngest known horizons of which are Awamoan in age.

Micaceous sandstones similar to those of Waipuna Bay are well exposed below scrub and on slipped ground on the east side of School Creek, and again above the type occurrence of Dowling Bay Limestone in the north-west corner of Dowling Bay. In both these cases, as in Waipuna Bay, the sandstone is interbedded with light-coloured, deeply weathered tuffs in its upper members and is overlain by basalt. Furthermore, jointing and flow structures in the basalt near School Creek and above Waipuna Bay dip northwards at 30° to 50° and are conformable with the bedding of the sediments. Likewise the flow structures in phonolitoid rocks above the basalt (fig. 2) also dip northwards. This does not of itself prove a conformable relationship. *Interbedding* of tuffs and sandstones however shows that the early pyroclastics are conformable with the sandstones.

At the north-west corner of Dowling Bay (S164-2/283819 (1942)) an apparent thickness of over 50 ft of Dowling Bay Limestone is exposed but minor faulting makes it difficult to measure a detailed section. Bands up to several feet thick of dark fetid limestone and calcareous sandstone are interbedded with more micaceous and argillaceous carbonaceous layers. Weathered sandy layers resemble decalcified members at Waipuna Bay. The limestone is overlain by micaceous sandstones and tuffs up to the lowest exposure of basalt.

On the basis of our mapping, the Dowling Bay Limestone should be represented in the lower part of the Waipuna Bay section. As noted above, calcareous micaceous sandstones and sandy limestone do in fact occur at Waipuna Bay, but they appear to be thinner and less carbonaceous than on the west side of Dowling Bay. Layers of coarse shells comparable to those in the typical limestone were not found in Waipuna Bay, the shell bed there being of different lithology and probably a slightly higher horizon. Nevertheless, its age is little if any younger than that of the limestone (Appendix A). We suggest that the apparent absence of typical Dowling Bay Limestone from Waipuna Bay is due partly to decalcification at the latter locality, and partly to rapid facies changes in what was clearly a shallow-water environment.

Waipuna Bay Formation

The Dowling Bay Limestone is defined by Benson (1959) as "Greyish impure limestone with thin, irregular interbedded mudstone layers at Dowling Bay near mouth Otago Harbour . . ." This does not include the sandstones described above. We therefore propose the name Waipuna Bay Formation to include all the sediments as exposed in the Waipuna Bay section up to and including the highest sandy layer, and thus excluding the highest 15 ft of tuffs unless better exposures should subsequently show sandstone members among them. We interpret the Dowling Bay Limestone as a member, or lens, of this larger unit.

The clastic grains in the Waipuna Bay sandstones consist essentially of quartz, albite, and muscovite, together with epidote, chlorite, zircon, and tourmaline, minerals that could be derived directly or indirectly from the Otago schists. There are also composite granules of semi-schist, sometimes containing pumpellyite, and varying amounts of glauconite. Orthoclase or microcline is a rare constituent.* The tuff members contain a minor admixture of the above constituents.

Structure

Finlay and McDowall (1923) and Benson (MS map) show a fault trending east-north-east which they believe has let down the phonolitoid masses forming Taylor Point and Otofelo Point from a position several hundred feet higher in the volcanic sequence. In support of the fault hypothesis it might be argued that basaltic agglomerate occurs on the shore beneath the road at the main limestone locality, and trachyte occurs 200 yd to the east, both on the down-thrown side of the supposed fault.

For the following reasons, we do not consider it necessary to postulate such a fault:

(a) The flow structures of the Taylor Point mass are steeply dipping or nearly vertical (fig. 2) and are consistent with an intrusive origin for the mass.

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*Although minerals derived from the Otago schists are the dominant constituents of the East Otago Upper Cretaceous and Tertiary sediments, abundant microcline and/or orthoclase of plutonic origin has been observed in some specimens of Abbotsford Mudstone (Upper Cretaceous to Lower Eocene); in a quartz-microcline-albite-muscovite sand occurring as sandstone dykes and as matrix in a coarse volcanic breccia exposed on Dunedin's Northern Outlet (grid ref. S164/161773 (1943)); and in sandy carbonaceous shale associated with the Waitati lignite (S164/180858). Like the andalusite, kyanite and clear tourmaline in the Eocene Green Island Loose Sand (Hutton and Turner, 1936) this indicates that an important component of the sediments of the Dunedin district came from a high-grade metamorphic and granite-bearing terrain, possibly the Stewart Island mass and any eastern extension of this mass that may now be submerged.

(b) The Otafelo Point mass has a dyke-like form and trends north-westwards across the supposed fault. It is lighter in colour at its seaward end than on the inland side owing to a lower content of aegirine-augite, but it appears to be monolithic.

Basalts and tuffs capped with hornblende trachyandesite dip westwards at Pulling Point. From here, there is a more or less continuous swing in strike through about 40° at Dowling Bay to 117° at the east side of Waipuna Bay. The dip is consistently outwards, forming an arc centred approximately on Taylor Point, the pattern recalling the outward dips surrounding the Varleys Hill and Dicks Hill intrusions of Otago Peninsula (Benson, e.g., 1940, fig. 1, Section CC). The Taylor Point intrusion does not of itself appear to be large enough to have caused doming of the magnitude observed, but we suggest, nevertheless, that the doming was caused by intrusion. Possibly the trachytoid phonolite now exposed on Taylor Point was emplaced following caldera-like collapse. The basaltic agglomerate and trachyte on the shore below the Dowling Bay Limestone may have been emplaced within the collapsed area at this time. Alternatively, the agglomerate and trachyte could be interpreted as structurally underlying the Dowling Bay Limestone, but against this possibility is the fact that the agglomerate itself contains fragments of micaceous sandy limestone. The significance of an east-west line of limestone or limestone blocks in the agglomerate is obscure.

Tuff Members of the Waipuna Bay Formation

The lowest ash we have recorded is the fossiliferous bed 2 ft thick, and 41 ft above the main shell bed at Waipuna Bay. Over a dozen large unweathered feldspar fragments were examined from this tuff and all were found to be calcic plagioclase, typical grains being in the range, An_{47-61}. The great majority of lapilli (up to 5 mm diameter) contain scattered tiny microlites of andesine-labradorite in a completely argillised matrix. A few lapilli were noted with trachytic texture. Most of these appear to be weathered trachyandesite, but some are trachyte. The tuff is interpreted as being of basaltic to trachyandesite composition with a minor admixture of trachytic fragments.

In a tuff 14 ft thick, 60 ft above the shell bed, zoned feldspars ranging from labradorite to oligoclase with anorthoclase margins were noted, together with a few large crystals of anorthoclase. A few of the lapilli in this bed are microcrystalline anorthoclase trachyte, but most lack microlites or phenocrysts, are completely argillised and were presumably originally quite glassy. It is thus not clear whether the tuff is essentially tracyhtic or basaltic although a trachytic component is certainly present. The tuffs and breccias making up most of the upper 73 ft of the sedimentary succession however are dominantly basaltic in origin in spite of their rather light weathering colours.

THE BLANKET BAY INLIER (FIG. 5)

On the south side of Blanket Bay, Otago Harbour, sandstones (S164/224765 (1943)) form a prominent cliff (Marshall, 1906, p. 389). They strike 146° and dip 37° south-westwards, and are overlain without visible discordance by flow rocks and sandy tuffs. The measured section is as follows:

H		Thick feldspar basalt, deeply weathered.
G	1 ft	Sandy tuff, coarse near base.
F	12 ft	Nephelinitoid phonolite, deeply weathered.
E	3 ft	Very deeply weathered flow rock, flow structures parallel to the bedding of the underlying sandstones.
D	45 ft	Well sorted, poorly bedded, white, quartz-albite-muscovite sandstone.
C	29 ft	Well bedded micaceous sandstone with muddy carbonaceous streaks and scattered waterworn pebbles of coal.
B	36 ft	Covered by road. Towards the north end of the cliff where the new main Port Chalmers Highway leaves the old, the following beds from this part of the section are exposed in the cliff on the east (upthrown) side of a small fault striking 113°: (c) Muddy micaceous sandstone as above. (b) Two feet of chocolate-brown and light grey trachyte tuffs with large anorthoclase crystals $(Or_{28}\ Ab + An_{72})$, coarse lapilli and sandy layers. Grades up into (c). (a) Two feet of yellowish brown basaltic tuffs containing copious crystals of sodic labradorite.
A	37 ft	Highly calcareous muddy sandstone incompletely exposed at low tide on the shore platform (? Oyster Point locality of Hector and Skey, 1866). It has yielded a poorly preserved foraminiferal fauna (S164/532) including *Textularia gladizea* Finlay, *Notorotalia wilsoni* Hornibrook, *Loxostomum truncatum ?* and *Notorotalia* n.sp. as in the Southland Series at Clifden, consistent with an Upper Southland Lillburnian or more probably Waiauan age (Mr Hornibrook, pers. comm.). It also contains solid siliceous spicules.

At Curles Point, several chains to the south, the upper well-sorted sandstone is again exposed, both on the promontory and on the tide platform on either side of the railway causeway. Here it is overlain by auguite basalt, followed by plagioclase basalt and then trachytoid phonolite quite distinct from the nepheline-rich rock already mentioned. Clearly some at least of the individual flows at this locality were of limited lateral extent.

Thin lenses of carbonaceous, micaceous sandstones are exposed at low tide on a minor promontory (S164/227769 (1943)) near the middle of Blanket Bay, where they are interbedded with trachytic and some basaltic tuffs and breccias and contain occasional leaf impressions. Their general strike is 61°, dip 27° south-east, but at the east end of the outcrop an attitude of 151°, dip 33° north-east was measured. In the base of the low cliff forming the promontory a patch of sandstone several yards long is exposed. It contains copious small carbonaceous markings and scattered trachytic fragments up to about 1 in. diameter, especially at the top where it is overlain by trachytic tuffs or breccias, passing into flow trachyte some yards to the west. Highly variably attitudes of the bedding within this one small exposure suggests that it is a raft, itself containing lapilli from an early eruption, which has been rucked up by an advancing trachyte flow. About a chain to the west, another disturbed block of dark, highly carbonaceous mudstone has its bedding strongly twisted in similar fashion. We interpret these sandstone lenses as being virtually contemporaneous with the trachyte, which presumably did not

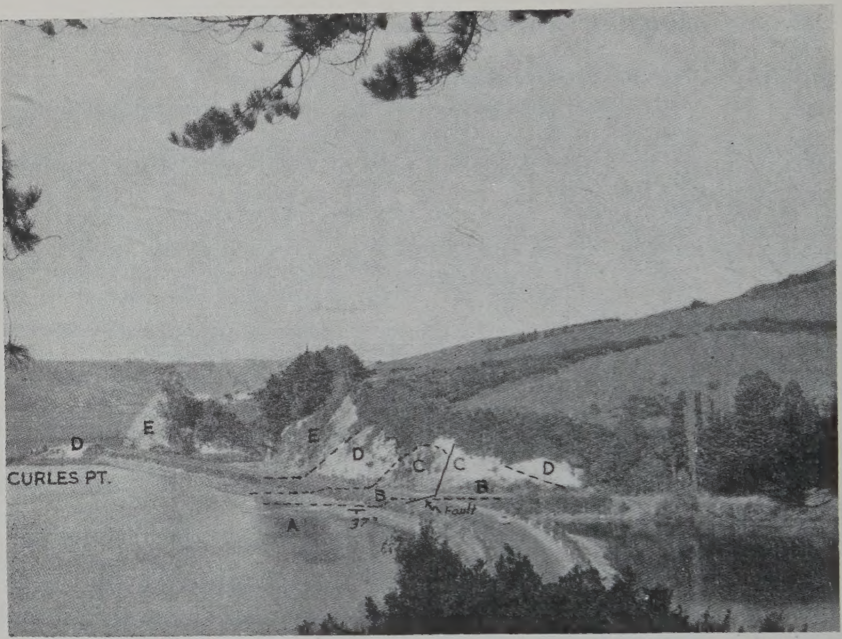

FIG. 5—Section of Blanket Bay, Otago Harbour. Volcanic rocks (E) overlie Waiauan sediments (D, C, B, and A – see text).

extend to the south end of the Bay where it is probably represented by the trachytic tuffs referred to above.

Further exposures of sandstones occur on the steep bank near the centre of Blanket Bay. Relationships are very confused. Individual outcrops of sandstone have variable dip, appear to underlie or, in places less decisively, to overlie trachyte, or they may be cut intrusively by trachyte. One such outcrop (S164/533) occurring near the highest point on the road about 100 ft above sea level, contains an abundance of siliceous spicules and spherules similar to those noted in (S164/532) and in some phases of the Dowling Bay Limestone. It is not clear to what extent these sandstone masses owe their present position to rafting by intrusive trachyte, to tectonic movement, or to post-trachyte deposition, although the suggested correlation of the trachytes with the tuff in the south end of the Bay would imply a considerable amount of local post-trachyte sedimentation.

CORRELATION OF WAIPUNA BAY AND BLANKET BAY TUFFS AND SANDSTONES

As demonstrated above, the Waipuna Bay Formation with its included tuffs is probably entirely of Waiauan or at youngest, early Tongaporutuan age at its type locality, as is the calcareous sandstone at Blanket Bay. Except perhaps for the upper clean sands, the lithology of the Blanket Bay sediments closely resembles that of typical members of the Waipuna Bay Formation with which we correlate them.

The correlation of the tuffs with Benson's volcanic sequence remains to be considered. Tuffs occur through at least the upper 120 ft of the Waipuna Bay Formation at Waipuna Bay. The characteristic trachytes of Benson's Initial Eruptive Phase, well developed at the base of the volcanic sequence near Port Chalmers and North Head, are entirely missing from the Waipuna Bay inlier, except for the small occurrence on the shore of Dowling Bay which we suspect to be intrusive, and except for the trachytic component of the tuffs as described above.

As noted above also, trachyte emplaced among sandstone and basaltic tuff at the centre of Blanket Bay is not continuous to the south side of the Bay where it appears to be represented by trachyte tuff overlying basaltic tuffs about 90 ft below the top of the sandstones. Benson (MS) has postulated that the Initial Phase trachytes were the differentiation product of a basaltic magma, no other surface manifestation of which was recognised apart from basaltic fragments in certain of the trachytic breccias. We suggest that the earlier tuffs in the Waipuna Bay Formation at both localities can be correlated with the Initial Eruptive Phase, the basaltic members being products of the basic magma postulated by Benson.

The uppermost basaltic tuffs and breccias below flow-basalt at Waipuna Bay can tentatively be ascribed to the early part of the First Major Eruptive Phase.

APPENDIX A

Molluscan Determinations from the Waipuna Bay Collections

By Anne Scott, New Zealand Geological Survey, Department of Scientific and Industrial Research, Lower Hutt

S164/524. Faunal list for shell bed near foot of old cliff, 75 yd due north and slightly east of north from Abernethys Creek culvert, Waipuna Bay, Grid ref. S164–2/294824 (1942).

Nucula cf. *crepida* Marwick.
Nuculana (Saccella) sp.
Nuculana (Jupiteria) sp.
Anomia sp.
Modiolus ? altijugatus Marwick.
? Limopsis sp.
Ostrea sp.
Atrina sp.
? Spissatella sp.
Pteromyrtea sp.
Angulus sp.
Bassina speighti (Suter).
? Kuia or *Marama* (inflated Venerid casts).
Eumarcia (Atamarcia) sp. (small).
Hedecardium sp. (fragment)
? Nemocardium sp.
Gari sp.
Panope sp.
Sigapatella sp.
? Maoricrypta (large, coarsely ribbed)
Struthiolaria (cf. from Waiauan of Lower Waipara and Weka Pass).
Willungia sp.
? Buccinulid fragment.
Alcithoe cf. *solida* Marwick.
Dentalium sp.
Brachiopods.
Coral (decorticated).

Field relations indicate that the collection is slightly younger than the Dowling Bay Limestone (Waiauan) with the fauna of which (Finlay and McDowall, 1923) the assemblage has much in common. The *Struthiolaria* is unlike the distinctive *S. praenuntia* (Marwick) of the Upper Tongaporutuan and *Spissatella* (doubtful identification) is not found beyond lower Tongaporutuan and seldom above Waiauan. *Willungia* is an Eratoid genus, not yet found in rocks younger than the Southland series, but it is a rare group. The fauna appears to be Waiauan or at the latest lower Tongaporutuan.

S164/524. Shell lens in 2 ft bed of tuff, 41 ft above S164/522. Grid ref. 294824. There are numerous small and featureless lamellibranchs which give no indication of age. There are some quite good Turritellids but the only other potentially useful fossil is a *Struthiolaria* fragment similar to Southland forms rather than Tongaporutuan, as far as can be judged from so small a piece.

S164/523. Micaceous sandstone in cutting under scrub-covered bank on road side, immediately west of Taylor Point phonolite quarry. Grid ref 288822. *Chlamys* sp., a form which turns up spasmodically from the Oligocene onwards.

S164/525. Fine buff sandstone from the bottom of the column, near base of old cliff, 298° from Abernethy's Creek culvert, Waipuna Bay. Grid ref. 293824. This contains good casts of *Neilo sublaevis* Marwick, a distinctive species of the Southland Series which occurs in the Ihungia and lower Tutamoes (Southland Series) of Gisborne and also in the Tongaporutuan of Taranaki. It has not been found in the Pareora Series.

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